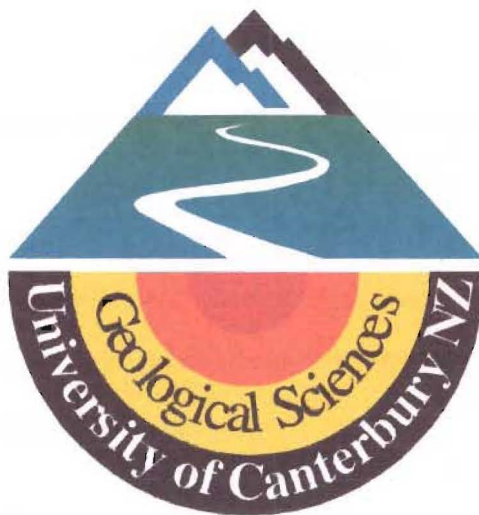


THE GEOLOGY AND GEOMORPHOLOGY OF THE CORONET PEAK AND ARTHURS POINT LANDSLIDE COMPLEXES.

A thesis
submitted in partial fulfilment
of the requirements for the degree
of
Master of Science in Engineering Geology
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ABSTRACT

The Coronet Peak and Arthurs Point Landslides have both formed in schist bedrock on the northwestern side of the Wakatipu Basin. The rock masses at both sites show evidence for the presence of many defects, such as foliation, schistosity, foliation shears, sheared and crushed zones, faults and joints, all which have formed as a result of the deformation and metamorphism the schist has undergone.

The Arthurs Point Landslide has a volume of $2.4 \times 10^7 \text{m}^3$, and is a wedge failure, with sliding occurring along the intersection of two master joint sets rather than foliation, the latter dipping into the slope. It is a retrogressive failure, with the eastern part having failed as a consequence of undercutting of the slope during the Last (Otiran) Glaciation. Actual failure of the eastern segment occurred after retreat of the ice left the valley sides over steepened and unsupported, whilst failure of the western part was related to the rock above the initial failure being left unsupported and incision of the Shotover River.

The Coronet Peak landslide is a translational planar slide with a volume of approximately $1 \times 10^9 \text{m}^3$ that failed along foliation and/or foliation shears, and the slide can be divided into zones (Zone A and B), based on morphological evidence.

Zone A is the larger zone (it has a volume of $6 \times 10^8 \text{m}^3$), and has well-developed hummocky topography, indicating continued but slow movement (probably less than 2-5mm/yr) as it has no form of toe support removal.

Initial failure of the Coronet Peak landslide was triggered by ice undercutting and subsequent retreat of the Wakatipu Basin, most probably during the Waimean Glaciation (approx 135 000 years ago), as it was during this ice advance that the Wakatipu Glacier extended all the way into the Wakatipu Basin and partially up the Shotover and Arrow Valleys.

Zone B can be divided into three subzones, which are all different reactivations. Zone B1 is a reactivated part of the Coronet Peak landslide and was triggered by the ice erosion related to the Otiran Glaciation. Evidence for this is seen on the valley floor at the toe of Coronet Peak landslide, with the Old Ben Lomond moraine situated adjacent to the end of the reactivated part. Most surficial glacial deposits in the Wakatipu Basin are related to the Otiran advance. Zone B2 is a reactivation within this, and was triggered by incision of the Shotover River. Zone B3 consists of two reactivations, both possibly related to flood events of the Shotover River.

Back analysis of the Coronet Peak Landslide were performed, to ascertain whether the model used for the landslide is appropriate. The results showed that the model is a reasonable one.

It is feasible to build on the Arthurs Point and Coronet Peak landslides, as long as it is not at the usual densities. With landslide debris, the foundation materials are not always compacted well, and therefore injudicious excavation could cause reactivations, like at the Coronet Plaza, where trimming a small amount of rock off the toe of the CPL to allow room for extensions caused some of the toe area to reactivate. The other important factor when building on landslides is surface and subsurface water, as more water in the slope than normal will increase the instability.

Building at the usual densities would increase the area being excavated, and so the disruption to the foundation materials, and it would increase the amount of water to be controlled.

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CHAPTER ONE

Introduction

1.1 Background Statement

The Coronet Peak Landslide and Arthurs Point Landslide complexes are two large, very slow-moving slope failures formed in schist bedrock near Queenstown, Central Otago. Coronet Peak Landslide is a translational failure controlled by foliation or foliation shears, while Arthurs Point is interpreted as a wedge failure controlled by the intersection of joint sets. The slides are heavily influenced by the defects in the schist bedrock, which formed during regional metamorphism and associated deformation.

Both landslides have been considered dormant, and no damage has occurred to any structures on the slides. However, it is more likely that the slides are creeping slowly, especially in reactivated areas, which can be identified on geomorphic evidence.

Residential development of the Queenstown area has occurred rapidly in the last twenty years, with the completion of more than twenty subdivisions over this time. Gradual infilling of the feasible building sites has increased pressure for building on marginal land. 'Marginal' land can be defined as "a building site affected by one or more geotechnical hazards with a frequency or severity such that during its design life damage may occur to the structure and/or injury may result to it's occupants" (Bell 1996). Therefore, Coronet Peak and Arthurs Point Landslides can be considered marginal land in terms of potential residential expansion.

The primary aim of this thesis is to map and interpret both failures, then to assess whether it is feasible to develop any sites on these landslides, and if so, to what densities. A broad zoning is also produced as a guide to further residential development.

1.2 Thesis Objectives

The principal objectives of this thesis are:

- a) To provide engineering geological maps at a scale of 1: 10 000 for the Coronet Peak Landslide and Arthur's Point Landslide complexes, and an overall geological map of the north-western part of the Wakatipu Basin at a scale of 1: 25 000.
- b) To assess and discuss the age and geological development of the Coronet Peak and Arthur's Point Landslide complexes.
- c) To interpret the failure mechanisms for both the landslide complexes and to assess the long-term stability implications for both slides.
- d) To assess the suitability of both the landslide complexes for expanded residential development.

1.3 Description of the Study Area

1.3.1. *Location of Study Area*

The Coronet Peak Landslide is situated on the south-facing side of Coronet Peak, and extends from across from Arthur's Point through Mt Dewar and Skippers Saddle to Coronet Peak itself, and down towards Arrowtown. Arthur's Point Landslide has formed on the northern face of Bowen Peak, which forms the southwest boundary of the Wakatipu Basin (Fig 1.1).

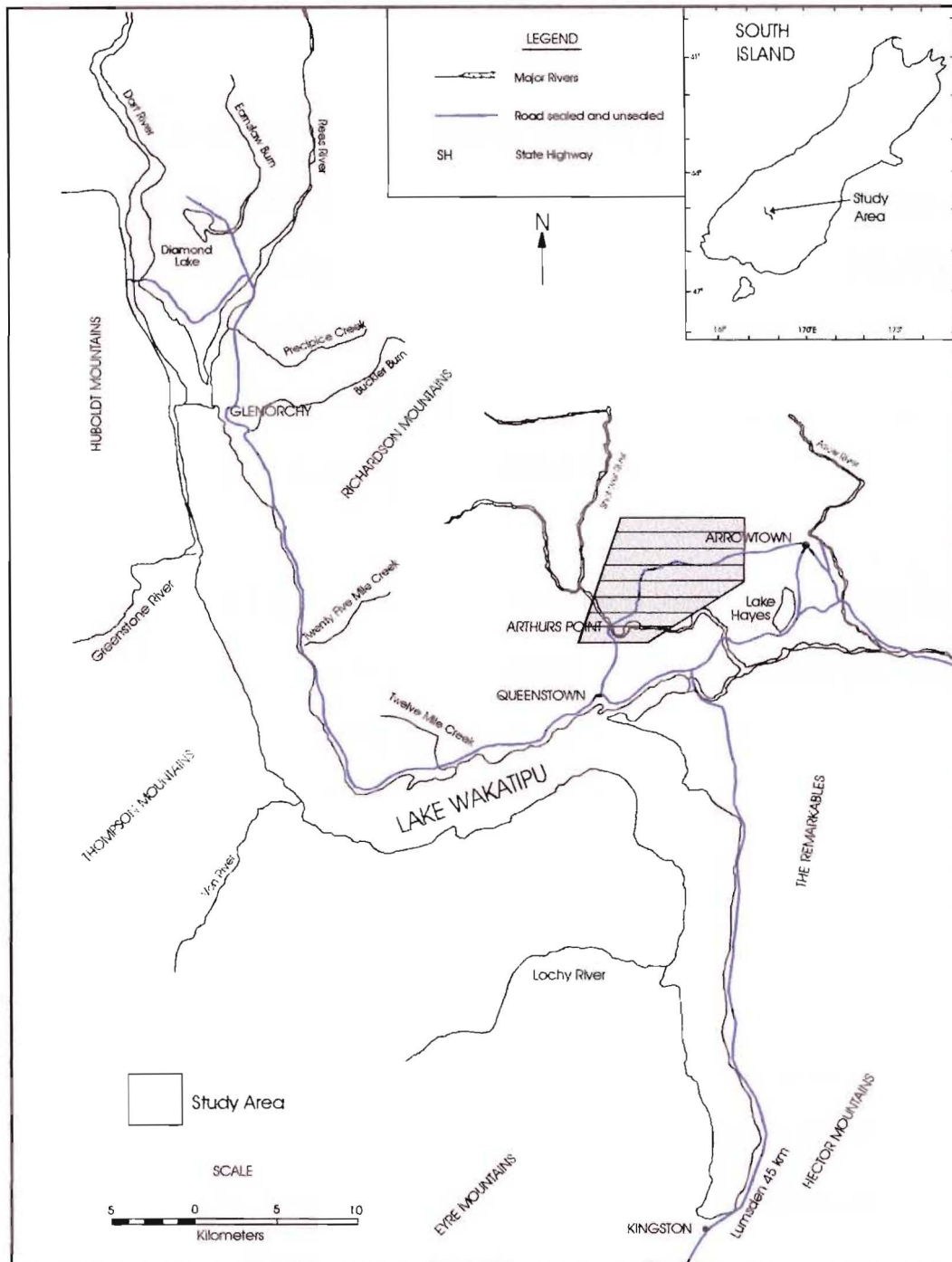


Fig 1.1: Location of study area. Study area marked by lined box.
(Base map from Watts 1988)

1.3.2 Topography

The highest topographic point in the study area is Coronet Peak at an elevation of 1651m, while the lowest point is along the Shotover River at an elevation of 360m. Coronet Peak Landslide has moderately steep slopes (15-25°), generally subdued scarps and deeply incised streams. The Arthurs Point Landslide typically has prominent scarps (15m+) and is steeper than Coronet Peak Landslide (35-45°).

Both slides have “ripple-landscape” (also called “sagging” metamorphic terrain), which is characteristic of ranges that have experienced recent uplift and/or glaciation (of which the Central Otago area has experienced both).

1.3.3 Existing Landuse

The land that makes up the Coronet Peak Landslide is used predominantly for pastoral activities, mostly grazing sheep with some farm-related buildings. The other main use is for recreation, including walking, mountain biking, hang-gliding and paragliding. Coronet Peak ski field, with its associated buildings, is located entirely on the Coronet Peak Landslide. Skippers Road starts at the toe of the landslide, just past the Coronet Plaza Hotel, and gives access to Skippers Canyon through Skippers Saddle.

Arthurs Point Landslide also has a predominantly pastoral landuse, mainly for grazing sheep. Part of the slide is residential, with 15 dwellings located on it. The Moonlight Track runs across the slide, and is used for walking and mountain biking. The Shotover Jet operates on the Shotover River below the landslide toe.

1.4 Regional Geologic and Geomorphological Setting

1.4.1 *Basement Geology*

In the Wakatipu Basin the basement rocks are entirely made up of Otago Schist (Fig 1.2), a part of the Haast Schist Terrane (Suggate 1961), which is a major metamorphic belt extending across the lower part of the South Island (Fig 1.3). The schist consists of greenschist, quartzofeldspathic and pelitic lithologies, and formed from an original sequence of sedimentary rock (sandstones and mudstones) and minor volcanic rock (Craw 1984).

The Otago Schist represents the deformed and metamorphosed amalgam of two greywacke terranes, the Torlesse Terrane to the north and east and the Caples Terrane to the south and west. These terranes are of different provenance, and their contact lies within the schist and is inferred mainly on the basis of geochemical discrimination between the psammitic samples (Mortimer 1993b).

Between the end of the Rangitata Orogeny (about 105 Ma) and the Mid-Tertiary (about 30Ma) the area that is now Otago was gradually worn flat by erosion, forming a broad, low relief landscape ('peneplain') (Barrell et al 1994). At about 40Ma the Moonlight Fault became active and movements on this fault caused localised subsidence of the earth's crust allowing a brief incursion of the sea from the west into the Wakatipu area. This sea, which had retreated by 15Ma, deposited a sequence of marine breccia-conglomerates, siltstones and bioclastic limestones (Turnbull et al 1975), which are known as the Bobs Cove Beds. Remnants are preserved as inliers along the Moonlight Fault at Bobs Cove, and as small isolated strips within the Shotover catchment to the north (Barrell et al 1994).

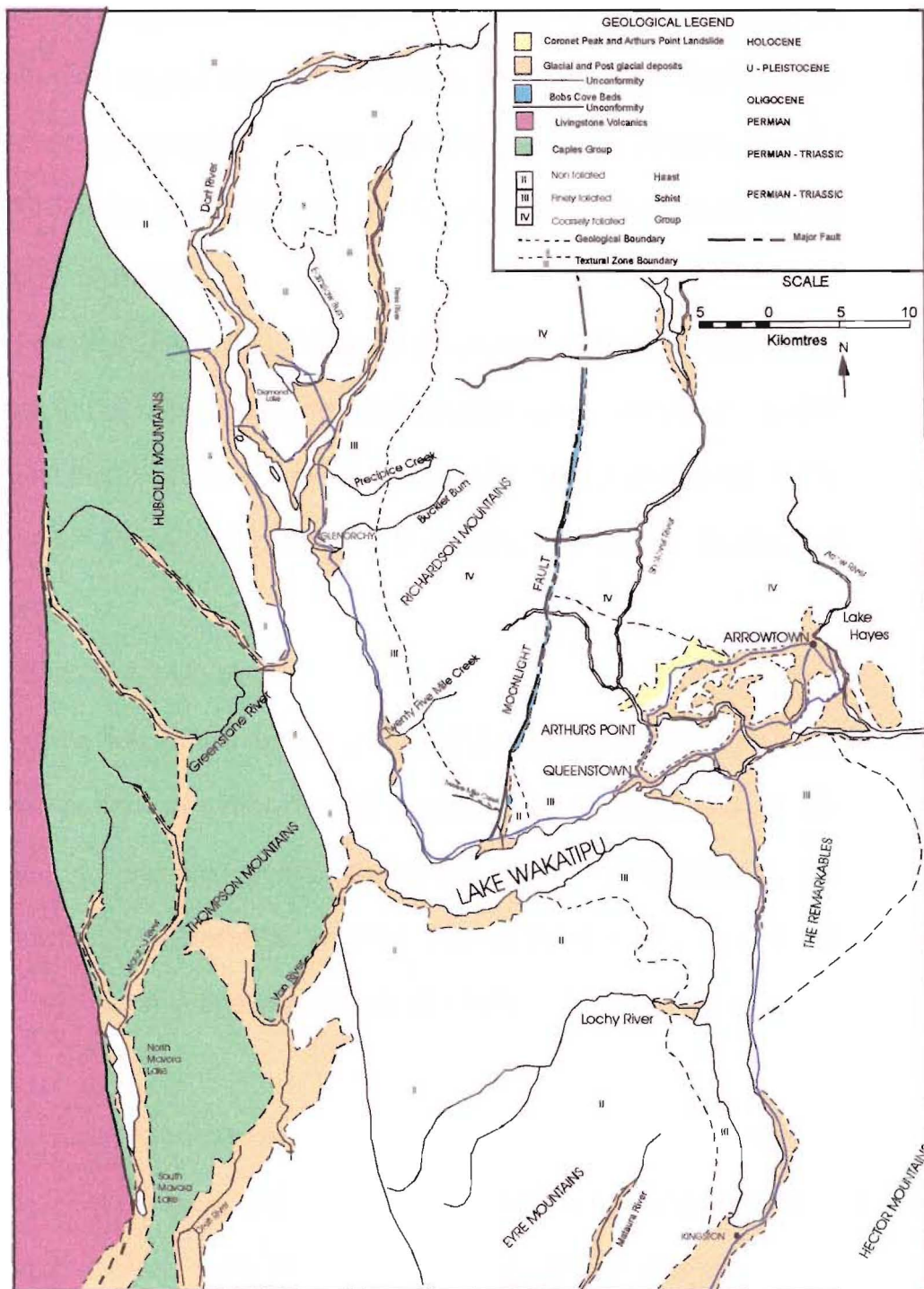


Fig 1.2: A Simplified Geological Map Of The Wakatipu Basin.
(Base map from Watts 1988)

In the later stages of the Tertiary Era (20-5Ma) broad subsidence of the Central Otago region caused streams to build up their beds, with the formation of freshwater lakes over the peneplain. Deposits from these lakes, streams and swamps occur at Coal Pit Saddle, Crown Saddle, and on the slopes above Gibbston Basin (Barrell et al 1994).

The Moonlight Fault (Fig 1.4) experienced two further periods of activity in the Tertiary, in the Late Oligocene and Miocene, while fault activity to the east did not start until the Pliocene (Turnbull and Forsyth 1988). Deposition of the Pliocene to Pleistocene Maori Bottom Gravels took place with the continuation of the Kaikoura Orogeny.

1.4.2 *Quaternary Geology*

During the last 2.5Ma drastic cooling of the earth's climate has occurred, fluctuating between periods of generally cold temperatures (glacials) and periods of relative warmth (inter-glacials). It is probable that at least 20 separate glacials have occurred during the 2.5Ma (Barrell et al 1994), though there is only evidence for four major glaciations in the Wakatipu catchment (Table 1.1) occurring during the last 500 000 years.

During glacial periods, summer temperatures were so low winter snow did not melt from the main ranges, and as a result extensive snowfields accumulated in the high precipitation areas along the Southern Alps and large glaciers spread down the valleys (Barrell et al 1994). As the ranges have risen, glacial action has carved out deep broadly U-shaped valleys, leaving eroded debris as moraines. Terminal moraine at Kingston is from the most recent glaciation (Otiran, 23 000 – 20 000yrs).

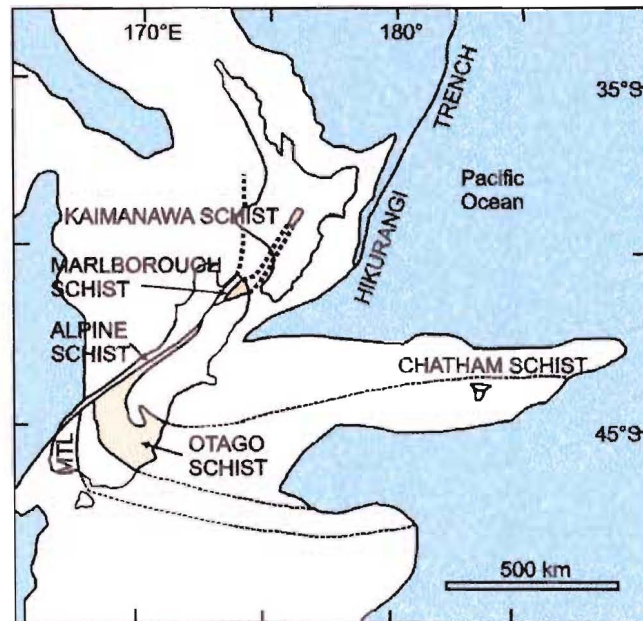


Figure 1.3: Distribution of the Otago, Alpine, Marlborough, Kaimanawa, and Chatham schists (Haast Schist Terrane) in relation to the Median Tectonic Line (MTL) and present-day plate boundary (heavy black lines). Offshore extent of New Zealand continental crust is left unshaded (Redrawn from Mortimer, 1993).

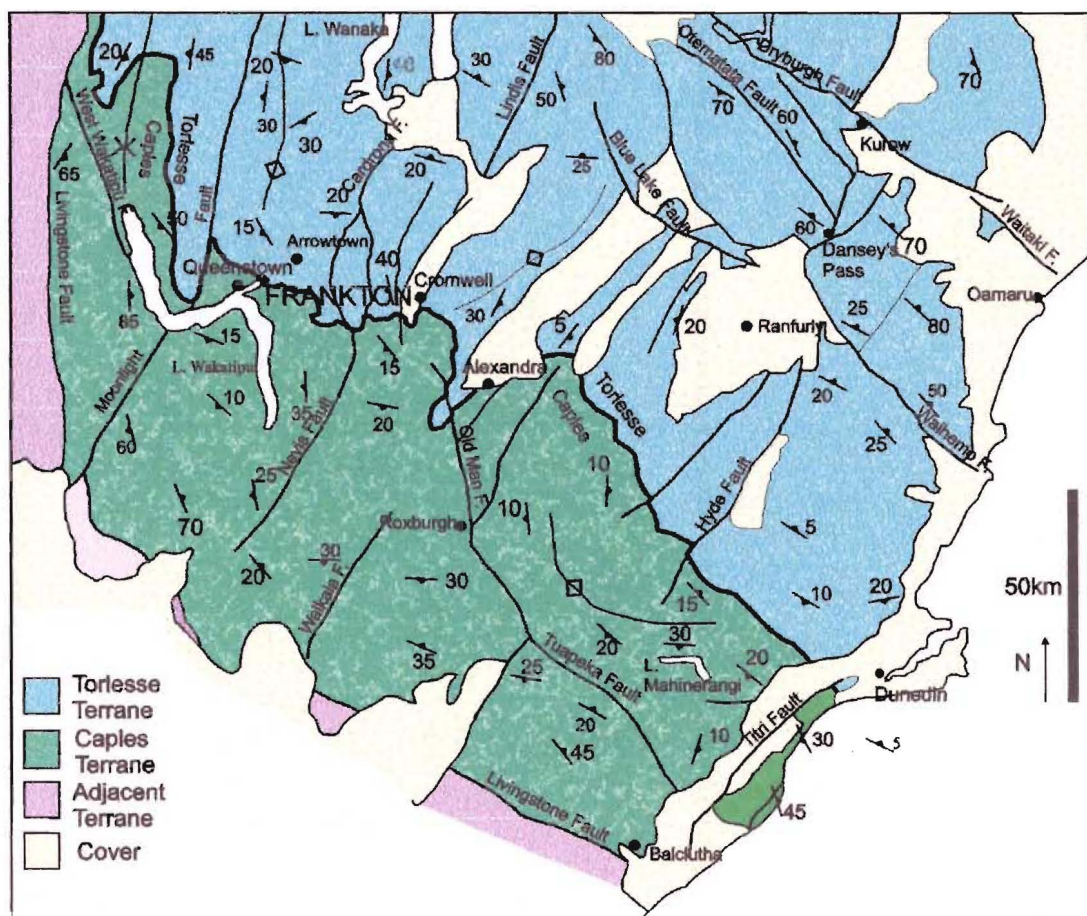


Figure 1.4: Simplified geological map of the Otago Schist and constituent terranes (Redrawn from Mortimer, 1993). Geology is from Mortimer (1992).

TABLE 1.1 : Sequence of Glacials and Interglacials, Westland and Wakatipu Basin							
<i>Oxygen Isotope Stage</i>	<i>AGE KY</i>	<i>Glaciation</i>	<i>Interglacial</i>	<i>Glacial Advance</i>	Westland <i>Glacial Formation</i>	<i>Shoreline Formation</i>	Wakatipu Basin <i>Glacial Formation</i>
1	0		Aranui	Kumara 3	Moana	Nine Mile	Arthurs Point/ Wye Creek Kingston (?)
2	12	Otiran		Kumara 2/2	Larrikins		
4	59			Kumara 2/1	Loopline		Kingston
5	71		Kaihinu			Awatuna Rutherglen	Oturian
6	128	Waimean		Kumara 1	Waimea		Garston2
7	186		Karoro			Karoro Scandinavia	Terangian
8	245	Waimungan		Hohonu	Tansey		Garston1
9	303						Waiwheran
10	339	Nemona			Cockeye		Athol

(Table from Suggate 1990, with Oxygen Isotope Ages from Imbrie et al 1984, and Wakatipu Basin Formation names from Bell 1983)

The Wakatipu Glacier carved out the Wakatipu Valley, and then split into two ice tongues when it met The Remarkables. The main ice tongue continued south towards Kingston, and at its greatest extent has reached south beyond Athol in the Mataura Valley (Fig 1.5). The other ice tongue extended northeast towards Arrowtown, carving out the Wakatipu Basin. In the largest recorded or preserved glaciation (the Waimean, from 160 000 to 110 000 years) the ice reached some distance down the Kawarau River, and subsidiary ice tongues pushed up the Arrow and Shotover valleys (Barrell et al 1994).

Climatic warming at the end of the last glaciation produced waning in the Wakatipu Glacier, which also caused the development of proto-Lake Wakatipu, dammed at Kingston by terminal moraine with discharge down the Mataura Valley (Watts 1988).

In the vicinity of Lake Wakatipu and Lake Hayes, delta, lake and beach deposits are extensively developed below about 400m elevation. The lake beach deposits provide clear evidence that Lake Wakatipu was formerly much higher than at present. Episodic lake lowering was achieved when a drainage outlet through the Kingston moraine was developed at about 400m elevation (Brockie 1973). Once the lake level reached an elevation of 355m there was a period of quiescence, which allowed extensive sedimentary deposits to build up at that level. The Kingston outlet was eventually abandoned as the Kawarau River outlet developed at a lower elevation. This was due to successive Late Pleistocene ice advances which progressively lowered the divide between The Remarkables and the Crown Range, allowing integration of drainage from the west of the Wakatipu area (Bell 1992). Major landsliding within the Wakatipu Basin and surrounding areas had already been initiated, generally before the onset of the Penultimate Glaciation (Waimean), producing an asymmetric profile in many glaciated valleys.

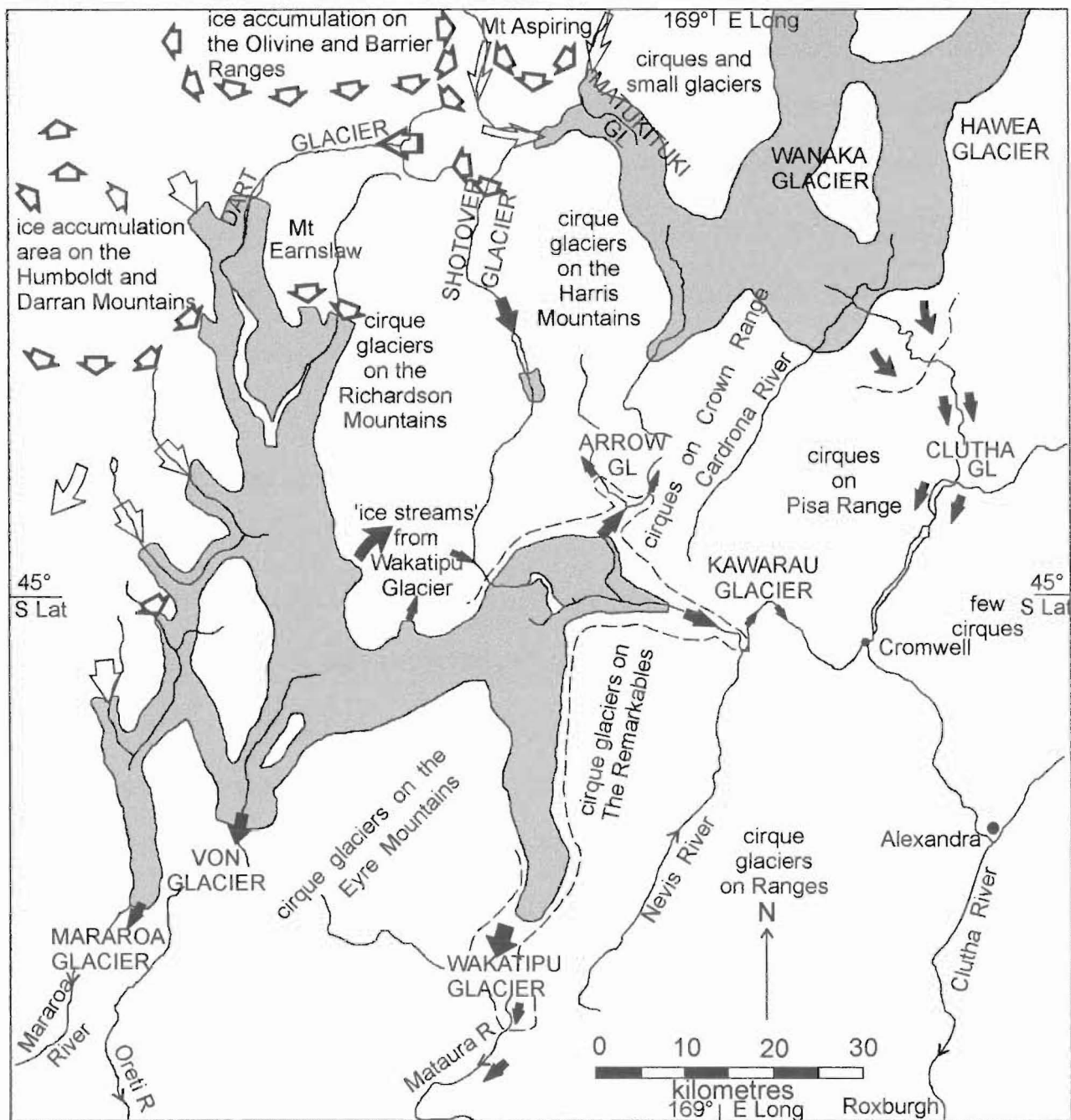


Figure 1.5: Map showing maximum ice limits for the Last (stippled) and Penultimate (dashed) Glaciations in the Wakatipu and Clutha catchments. Open arrows show the approximate directions of ice movement away from the snow accumulation areas, and the infilled arrows indicate the resultant glacier patterns (Bell, 1992).

1.5 Previous Investigations

1.5.1 *Geological Investigations*

Hector (1863) was one of the first people to produce a map of the Queenstown area, and recognise the Wakatipu Basin had been extensively glaciated. Another of the earlier investigations of the Wakatipu Basin was undertaken by Park (1908) and numerous other investigations of the area were done (1906, 1909) for the NZGS.

The Otago Schist was initially interpreted as a feature of the New Zealand Geosyncline, which formed during the Rangitata Orogeny (Fleming 1920, Reed 1958, Coombs 1960). It is now thought that the schist formed during the collision of the Torlesse and Caples terranes beneath a fore-arc region (Graham and Mortimer 1992).

The Caples terrane structure was studied by Turnbull (1980), and the Torlesse has been divided into three sub-terrane (Bradshaw 1989) called the Rakaia, Eskhead and Pahau.

The lithology of the schist is described by Craw (1984) and the chlorite zone was divided into four textural zones by Hutton and Turner (1936). The age of the onset of metamorphism is generally thought to be about 200Ma (Graham and Mortimer 1992), but ranges from 200Ma in the prehnite-pumpellyite facies rocks to 70Ma in the greenschist facies rocks.

The boundary between the terranes was decided by Mortimer and Roser (1992) on the basis of geochemical parameters to distinguish psammitic samples. They also concluded that the Alpine Lithologic Association (ALA), which was thought to be a separate terrane, is actually just a more segregated part of the textural zone IV of the Torlesse terrane. Lithologies from the ALA are geochemically and lithologically indistinguishable from similar lithologies in the Torlesse part of the Otago Schist.

Bishop (1972) identified three phases of folding and subdivided the chlorite 2 and 3 zones based on detailed mineralogy and metamorphic petrology, though he still thought in.

The first sediments deposited on the schist were the Henley Breccia, Shag Point Group and the Kyeburn Formation (Bishop and Laird 1976). These were a precursor to the Peneplain, which was first recognised by Cotton (1917). The Tertiary sediments laid down on the peneplain are only preserved in a few localities due to activity along several fault zones.

1.5.2 *Geomorphological Investigations*

Park (1909) systematically mapped foreign rock types in the catchments of the Wakatipu Basin, which provided the basis for the reconstruction of the geomorphic evolution of the area. This work was done before the concept of multiple glaciations had been suggested, and therefore Park thought in terms of a single ice shelf covering the Wakatipu Basin area in the Late Pleistocene times, with the higher mountain peaks projecting above it's surface.

Gage (1958) and other workers demonstrated multiple glaciations in New Zealand, while Suggate (1965) established a four-glaciation framework, which is used to interpret and correlate late Quaternary climatic events.

Garnier (1948) first described Central Otago as a basin and range province. He also identified the peneplain and described its development, and noted the tors. Garnier realised that Lake Wakatipu had been formed by glacial action, which left its terminal moraine at Kingston, damming the drainage outlet there and holding the lake at its present level. A partial chronology of geomorphic events in the Kawarau Valley is given by Bell (1992), summarising its evolution from Tertiary tectonism through to Late Quaternary glaciation and fluvial activity.

1.5.3 *Engineering Geological Investigations*

Limited engineering geological investigations have been carried out in Central Otago. In 1985 Bell and Pettinga produced a paper on engineering geology and subdivision planning. This paper (Bell and Pettinga 1985) used Fernhill as an example of how to undertake an engineering geological investigation of subdivisions, and provided a framework for all subsequent investigations.

Johnson (1986) mapped the Gibbston and Muddy Creek Landslides, dividing them into zones based on morphological and lithological differences, and assessed how inundation from the reservoir of a proposed hydroelectric dam in the Kawarau Valley would affect the stability of the slides.

The K9 Landslide has been mapped by Bell (1987), looking in detail at rock slope failure mechanisms. Among the mechanisms discussed were toe buckling and gravitational spreading.

An engineering geological investigation of many different deposits in the Wakatipu Basin was carried out by Watts (1988) to assess suitability for roading aggregate. Deposits that were considered good potential aggregate sources were looked at in detail to decide which would produce the highest quality roading aggregate. Six Late Quaternary sources were identified, and out of these a mixture of schist and greywacke was determined to be the highest quality roading aggregate source in the Wakatipu Basin.

In 1994 Cunningham looked at the engineering geological properties of the deposits in the Wakatipu Basin, and assessed the hazards present that would affect the Wakatipu Basin and the Queenstown Township itself.

Bell (1988 to 1997) and others investigated many subdivisions in Queenstown and the Wakatipu Basin, which resulted in a number of unpublished consultant reports. These reports give detailed engineering geological information about small specific sites, of which about fourteen relate to specific areas on the Coronet Peak and Arthurs Point Landslide complexes.

1.5.4 *Geotechnical Investigations*

A reasonable amount of geotechnical work has been done. All the engineering geological reports by Bell also cover geotechnical information. Moody (1985) carried out Point Load tests on intact quartzofeldspathic and micaceous schist samples from Maniototo, and found that all lithologies are stronger perpendicular to foliation. He also found densities for crushed zones, with the bulk density ranging from 2302 – 2051 kg/m³ while the dry density ranged from 1734 – 2063 kg/m³. Schmidt Hammer Rebound tests were also done, which gave values of 33 – 52 for the quartzofeldspathic schist, and 0 – 31 for micaceous schist. These values represent the mean rebound number, with the 0 value coming from pelitic schist rock material that disintegrated after 5 tests.

Johnson, in 1986, tested the strength of crushed schist from foliation shear zones, and found it had a residual strength $c_r = 0$ and $\phi_r = 27^\circ$ to $c_r = 0$ and $\phi_r = 32^\circ$. These values were obtained from performing a ring shear test, for the purpose of assessing the mobility of a soil at residual strength. He also found that the Plasticity Index of gouge from shear zones within the slides was 20 to 25. Using the Casagrande resistance envelope method, values were found of $c_r = 0$ and $\phi_r = 22^\circ$ to $c_r = 0$ and $\phi_r = 28^\circ$ for low normal stresses (<150 kPa). This method is a commonly used method of field analysis, and was used in this case to calculate the field strength of the slope debris because representative samples could not be tested in a laboratory.

Watts (1988) evaluated the geotechnical properties of glacial and fluvial deposits for roading aggregate. Grainsize analysis of the surficial deposits in the Wakatipu Basin was carried out by Cunningham (1994). She found that the tills had relatively low clay content (1 to 14 %). The clay content was lower in sandy tills and higher in silty gravel (5.6 to 39 %), which therefore had low cohesion.

A summary of geotechnical work done is given in Appendix Two.

1.6 Investigation Methods

1.6.1 Field Investigation

The bulk of the fieldwork was done from January 1988 to the end of February 1998, including detailed mapping of Coronet Peak Landslide for the 1:10 000 map (Fig 1.6a and 1.6b) and cross-sections (Fig 1.7). Additional mapping of Coronet Peak Landslide and limited sampling of sag pond and glacial material was conducted in May 1998. A final period of fieldwork was undertaken from January 1999 to February 1999 to map the Arthurs Point Landslide at a scale of 1:10 000 (Fig 1.6a).

1.6.2 Laboratory Studies

No laboratory testing was done of any samples. Many tests have already been performed by various people on schist throughout the Central Otago region. Tests have been done on intact greyschist (psammitic and pelitic) and greenschist, gouge from foliation shear zones and basal failure zones, and chaotic debris (landslide material). The schist in the Cromwell Gorge, Kawarau Valley, Gibbston Valley and Wakatipu Basin is all textural zone IV schist.

It was therefore considered that there was nothing to be gained by repeating tests that would have been relevant because the material that would have been tested is so similar to the schist actually tested.

The objectives of this thesis are not to test the schist anyway, but to analyse the landslides sensitivity to variable input parameters. A summary of what tests were performed and the results found is given in Appendix Two.

1.6.3 *Data Analysis*

Using data obtained already, factor of safety stability analyses were performed using the Galena program to confirm mechanisms that initiate failure. Back analyses were also performed

1.6.4 *Thesis Layout*

In Chapter Two an overall review of existing literature on landslides in schist terrain has been done, focussing specifically on the Cromwell Gorge landslides. After the building of the Clyde Dam, filling of Lake Dunstan could not be started until the slides were investigated in great detail. Filling of the lake would result in the inundation of the toes of most slides, so detailed investigation meant that the stability could be assessed using an appropriate model.

Chapter Three gives details on the major landslides in the NW part of the Wakatipu Basin. The extent, nature, failure mechanisms and present stability of each landslide are covered, as well as comparing the previous mapping done with the mapping done by the author.

Chapter Four presents the Coronet Peak Landslides movement history, from the age of the initial failure through to the most recent reactivations.

Chapter Five details the geomorphic history of the Coronet Peak Landslide and discusses the planning implications associated with residential development on both Arthurs Point and Coronet Peak Landslide complexes. The summary and conclusions are presented in Chapter Six.

CHAPTER TWO

Landslides in Schist Terrain

2.1 Introduction

Schists are metamorphic rocks having a pronounced anisotropy. This anisotropy means there is a pronounced weakness in the schist. Most landslides that develop in schist terrain usually exploit this weakness. Schist forms in areas of high tectonic activity usually associated with uplift, and glaciation. Retreat of the glaciers leaves the valley sides unsupported and subsequent incision by streams undermines the toe of the slope, increasing instability and eventually causing landsliding on slopes where the foliation dips into the valley. This has occurred in Italy, in the Pre-Alps located between Verbano and Cusido Lakes.

No matter where in the world schist rocks are found, they all have common properties. These include foliation or schistosity, and defects, such as foliation shears, crushed zones, joints and faults. These features form as a result of the many phases of faulting and folding that they experience.

The majority of the failures are dipslope failures, where planar failure has occurred on the foliation of the schist itself or the foliation shears, which are orientated downslope at an angle of approximately 25° (eg Downie Slide which borders the Revelstoke Dam reservoir, Canada).

Wedge failures also occur (eg Arthurs Point), due to planar failure along foliation or foliation shears with the intersection of defects such as faults and joints forming the wedge. Toppling failures are only likely where the angle of the slope is greater than 70° . Complex failures such as toe-buckling or sackung development are precursors to complete slope failure and are quite common.

When building engineering structures on or near landslides, the properties of the schist and landslide debris need to be taken into account. At Cromwell Gorge in New Zealand, intensive investigation was carried out to define the landslides (about 20 in number) that were to border the reservoir of the Clyde Dam. The filling of the lake caused the inundation of the toes of most of the landslides, and remedial and stabilisation works were carried out to reduce the effects of lake filling on the stability of the slides.

A similar situation occurred at Revelstoke Dam in British Columbia, Canada, except that the Downie Slide was the only landslide affected by reservoir filling (Fig2.1). In this chapter the Lake Dunstan landsliding project is covered in detail because it is the most extensive, in-depth investigation of schist landslides in New Zealand and perhaps the world. Other landslides from the Kawarau Gorge and Wakatipu Basin are also covered.

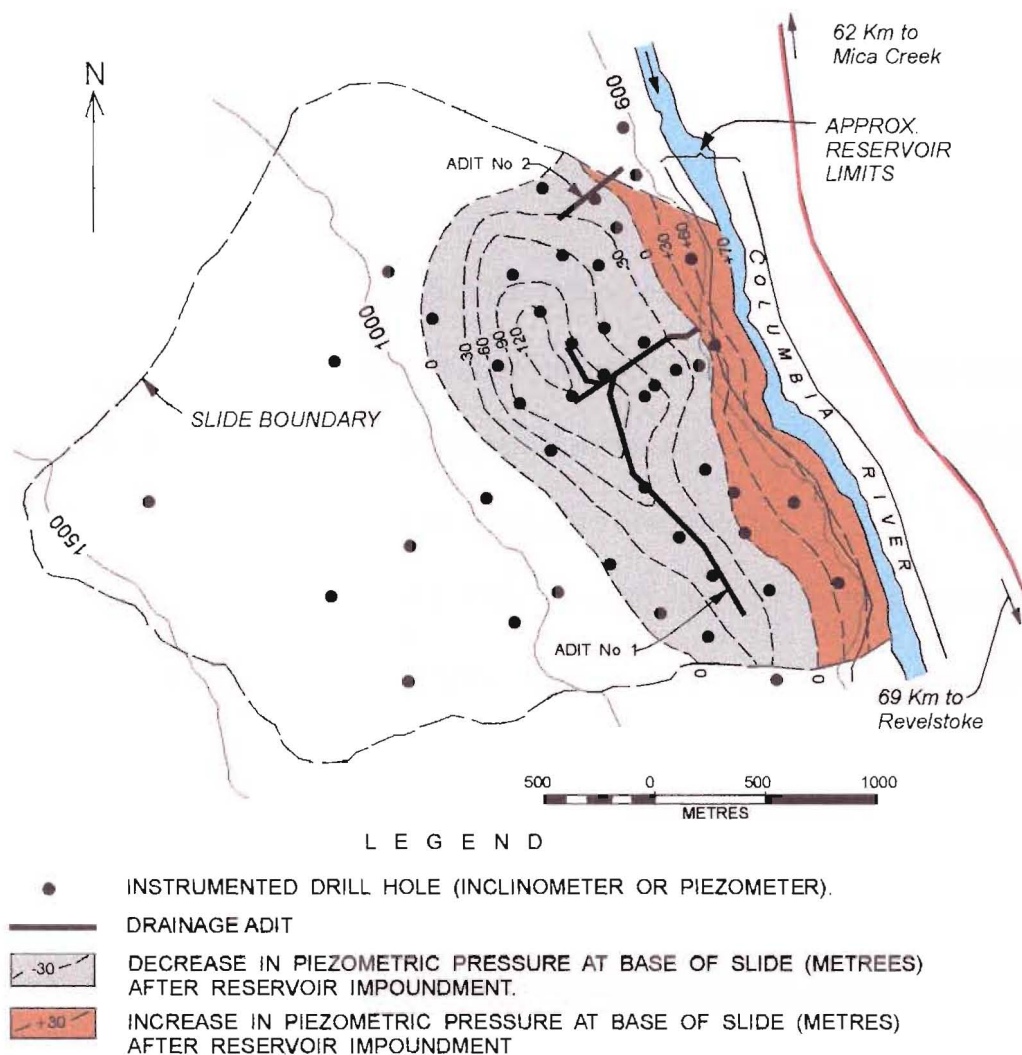


Fig 2.1 (a) : Plan of Downie Slide showing instrumentation array and drainage results. (From Imrie et al 1992).)

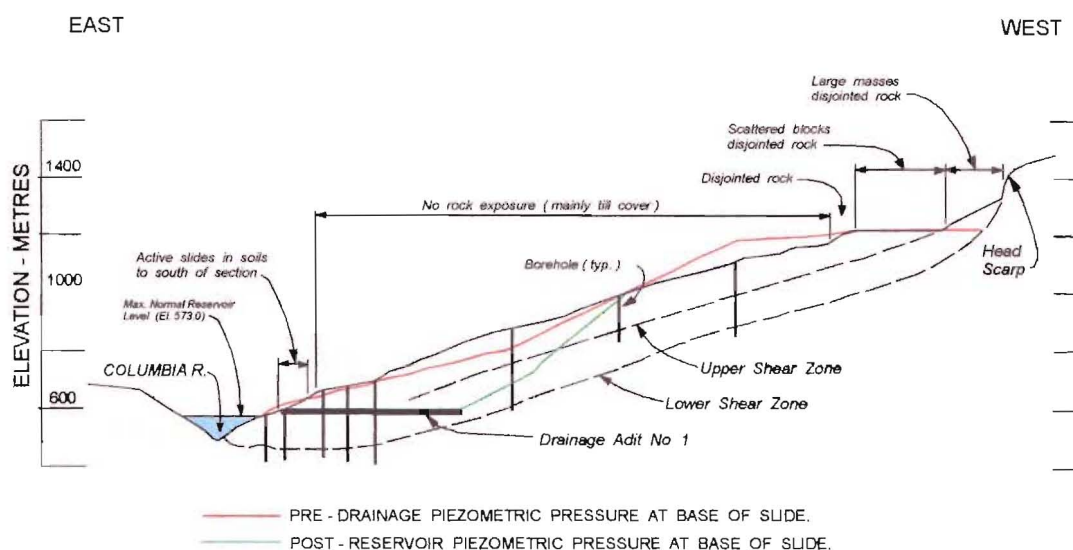


Fig 2.1 (b) : Cross-section of Downie Slide (From Imrie et al 1992).

2.2 Nature and Occurrence

2.2.1 Failure Mechanisms

2.2.1.1 Planar Failure

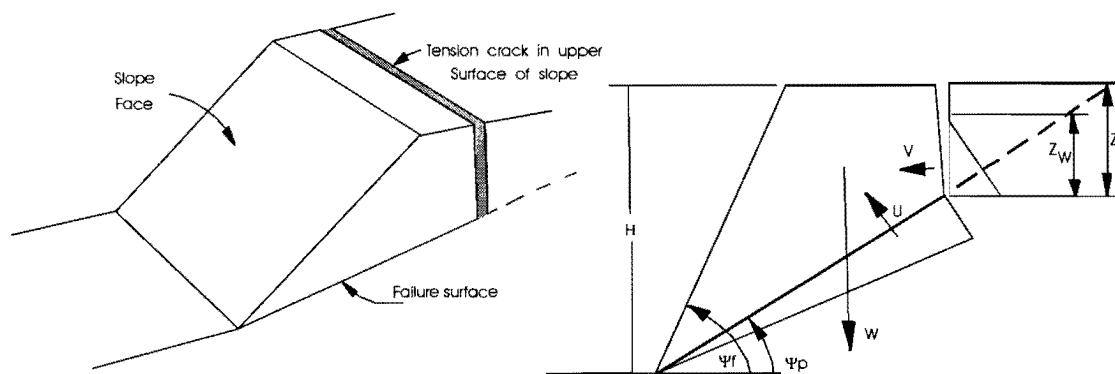
Planes of weakness may be found in the schist, especially in the phyllosilicate- (or mica-) rich layers. These include foliation, where the dark phyllosilicate minerals are segregated from the light quartzofeldspathic minerals into layers, and schistosity, where the minerals align themselves parallel to one another.

Both can have a significant effect on the strength characteristics of the schist, as it is usually significantly stronger when stress is applied perpendicular to foliation than when stress is applied parallel to foliation (MacFarlane et al 1992).

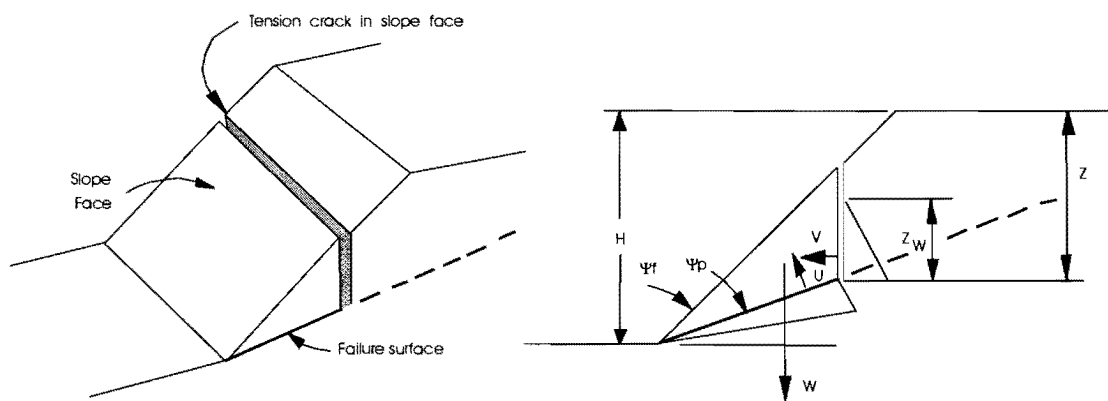
Therefore, there is a natural tendency for planar failure to occur along these weaknesses, especially when the foliation dips at an angle less than the slope angle and greater than the angle of friction. Usually failure propagates along defects such as foliation shears, schistosity and faults, and these defects generally dip between 20 and 30° (Fig 2.2). Planar sliding involves dipslope movement along the foliation or any defects sub-parallel to foliation (such as foliation shears, crushed zones, and can lead to marked valley asymmetry (eg Mt Malcolm, opposite Nevis Bluff).

2.2.1.2 Wedge Failure

Wedge failures occur where there is an increasing role of the prominent joint sets, which intersect with the schistosity or other joint sets (Fig 2.3). Fault influence can also occur, such as at the Gibbston Slide (Johnson 1986). Here, a dipslope/wedge failure has occurred between the Gibbston Fault and the schistosity with sliding happening on the fault crush zone surface.



Geometry of slope with tension crack in upper slope surface.



Geometry of slope with tension crack in slope face.

Fig 2.2: Three dimensional drawing of planar failure along a discontinuity that daylight in the slope face (From Hoek and Bray 1981)

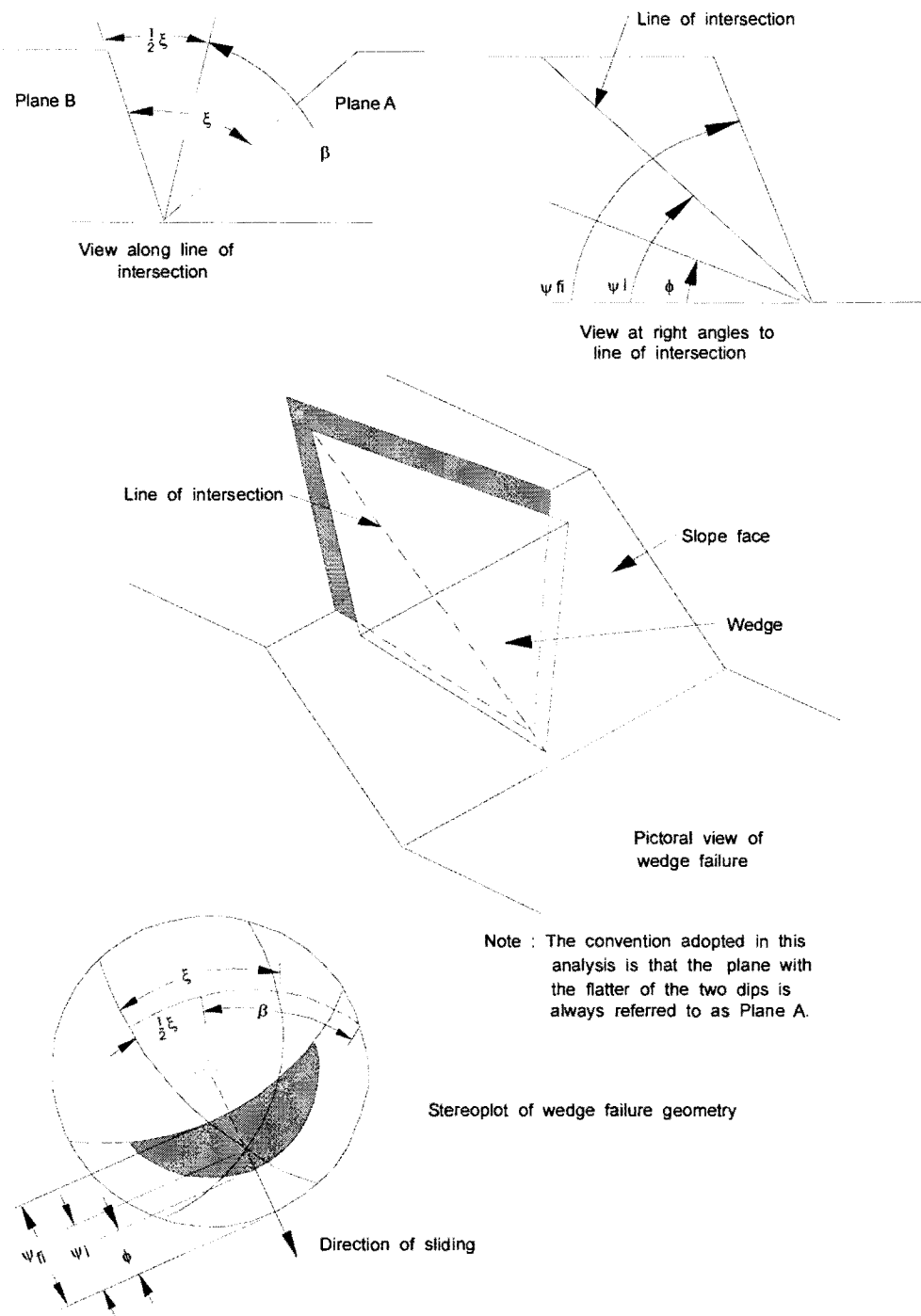


Figure 2.3 : Wedge failure geometry. (From Hoek and Bray 1981)

2.2.1.3 Complex Failure

Complex failures can also occur, such as toe buckling, gravitational spreading and toppling failures. Toppling typically occurs when defects dip into the slope face. Forward movement is by rotation (tilting without collapse), with shearing in the mid-section and simple slide failure at the base. Toppling only becomes a factor where the foliation dips at greater than 70 degrees (Fig 2.4)

2.2.1.4 Gravitational Spreading

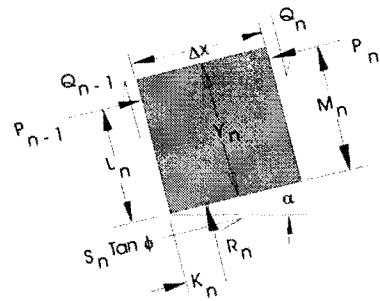
Gravitational spreading is recognised by topographic features, especially horizontal linear trenches and uphill facing (antislope) scarps on steep slopes ('sackung').

Radbruch-Hall et al (1976) describe sackung as "large-scale gravitational movement along a series of disconnected planes or by plastic deformation of a rock mass without formation of a through-going slide plane".

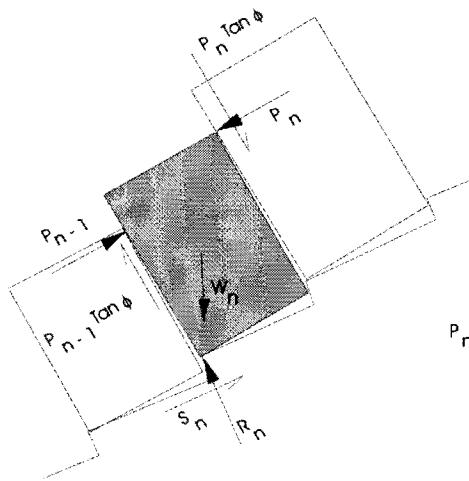
Bell (1987) observed that toe buckling and infilled topographic depressions above the K9 headscarp provided evidence for slope movements prior to and a consequence of large-scale failure. The evidence for toe buckling and gravitational spreading is summarized in Figure 2.5.

2.2.1.5 Toe Buckling

Toe buckling is a precursor to large-scale slope failure. Steepened to overturned (sub-vertical) schist results from toe buckling associated with creep deformation or slow downslope relaxation of the schist rock mass, consequent on any form of toe support removal. As the strength of the schist rock mass is less than intact rock, and its ductility is greater, it is apparent that the schist could respond by buckling (Johnson 1986).

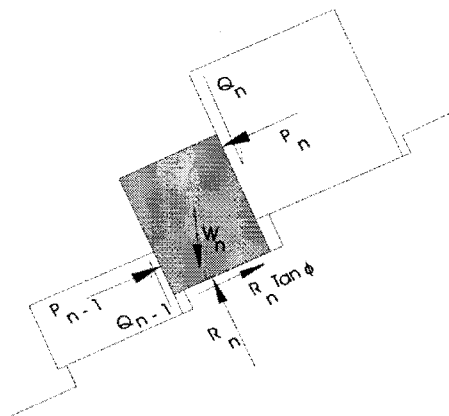


a) Forces acting on nth block



b) Toppling of nth block

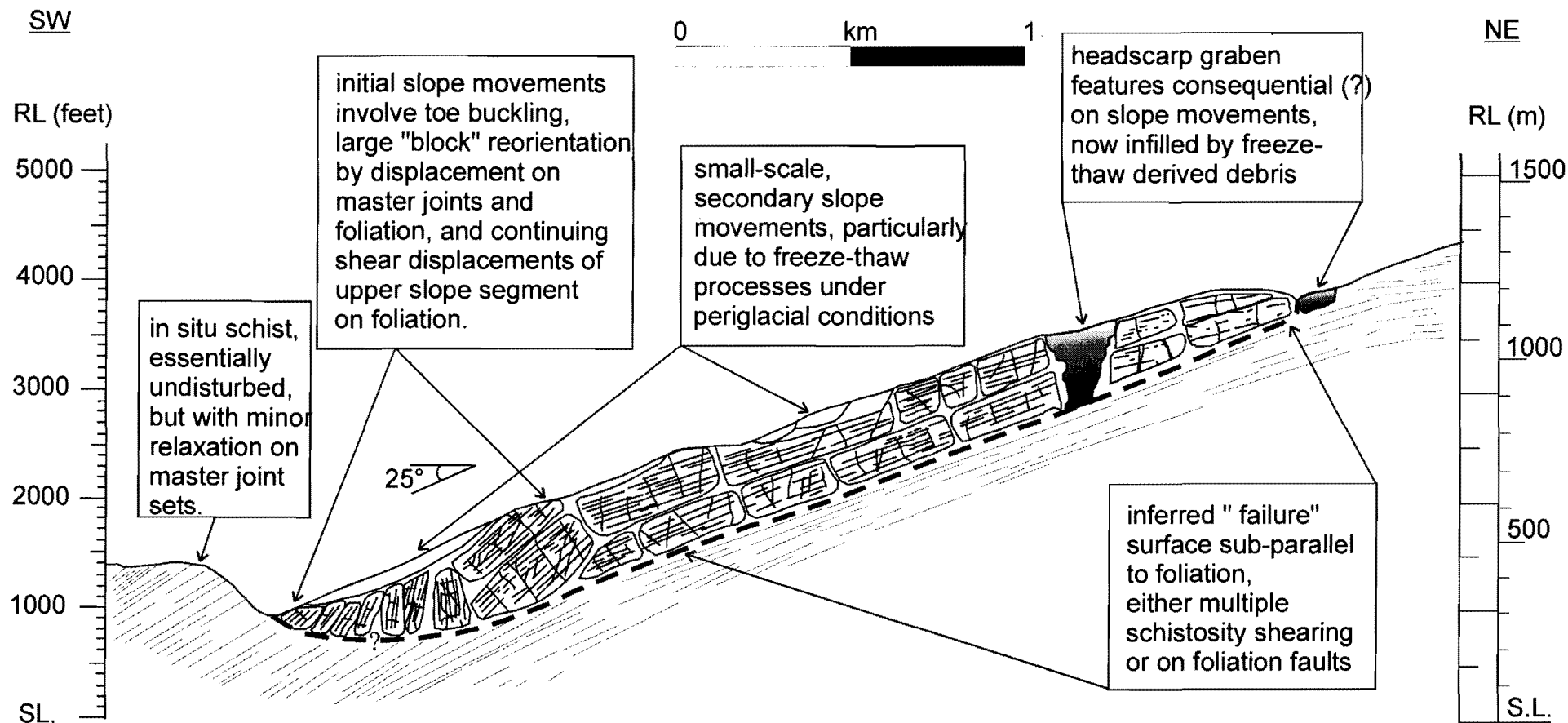
$$P_{n-1} = \frac{P_n (M_n - \Delta x \cdot \tan \phi) + (W_n / 2) (Y_n \cdot \sin \alpha - \Delta x \cdot \cos \alpha)}{L_n}$$



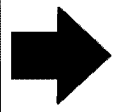
c) Sliding of nth block

$$P_{n-1} = P_n \frac{W_n (\tan \phi \cdot \cos \alpha - \sin \alpha)}{1 - \tan^2 \phi}$$

Fig 2.4 Limiting equilibrium conditions for toppling and for sliding of the nth block.



LATE QUATERNARY
TECTONISM AND
CLIMATE CHANGE



UPLIFT OF PISA
RANGE AND INCISION
OF LOWER KAWARAU
VALLEY



TOE BUCKLING AND
GRAVITATIONAL
"COLLAPSE" OF
SCHIST DIP-SLOPE



VALLEY AND SLOPE
MODIFICATION BY
FLUVIAL, GLACIAL AND
PERIGLACIAL PROCESSES

POSSIBLE CONTRIBUTING FACTORS

foliation attitude and geotechnical properties of schist;
creep on foliation shear zones; earthquake shaking due
to Pisa Fault movements; valley deepening by ice.



Figure 2.5 : K9 Landslide, Kawarau Valley, Central Otago, showing toe buckling and gravitational spreading. Note the obvious schist block rotation in the lower part of the section. (From Bell, 1983).

Such buckling deformation on a dipslope would be asymmetric due to the inclined anisotropy of the rock mass, and a fold would form. Gravity slope movements and confined groundwater (Fig 2.6) could also adversely affect this fold.

Toe buckling folds initially form by stress relief, but it is considered that they can develop further under gravity and groundwater pressures. Eventually this will reach the point where it will shear through the zone of maximum deformation. Subsequent downslope movement along this surface juxtaposes low angle schist above 'sub-vertical' schist, which are separated by a wide basal failure zone (Fig 2.6; Beetham et al 1992).

2.2.2 Failure Controls

2.2.2.1 Foliation Attitude

An important control on landslide distribution is foliation attitude, because most slides seem to develop on foliation dipslopes (where the foliation dips into the valley). Mica-rich lithologies (muddy, pelitic schists) tend to be weaker than quartz-rich layers, so there is a tendency for splitting along the mica layer (schistosity). Mica-rich layers are generally weak to moderately strong (σ_c 10 – 50 MPa). Thin gouge seams commonly control crushed and sheared zones. Gouge is the crushed u p clayey schist –derived infill in defects that have experienced movement. The gouge consists of moderately to highly plastic clay (30%), and the seams are from 0.5mm to 10's mm thick.

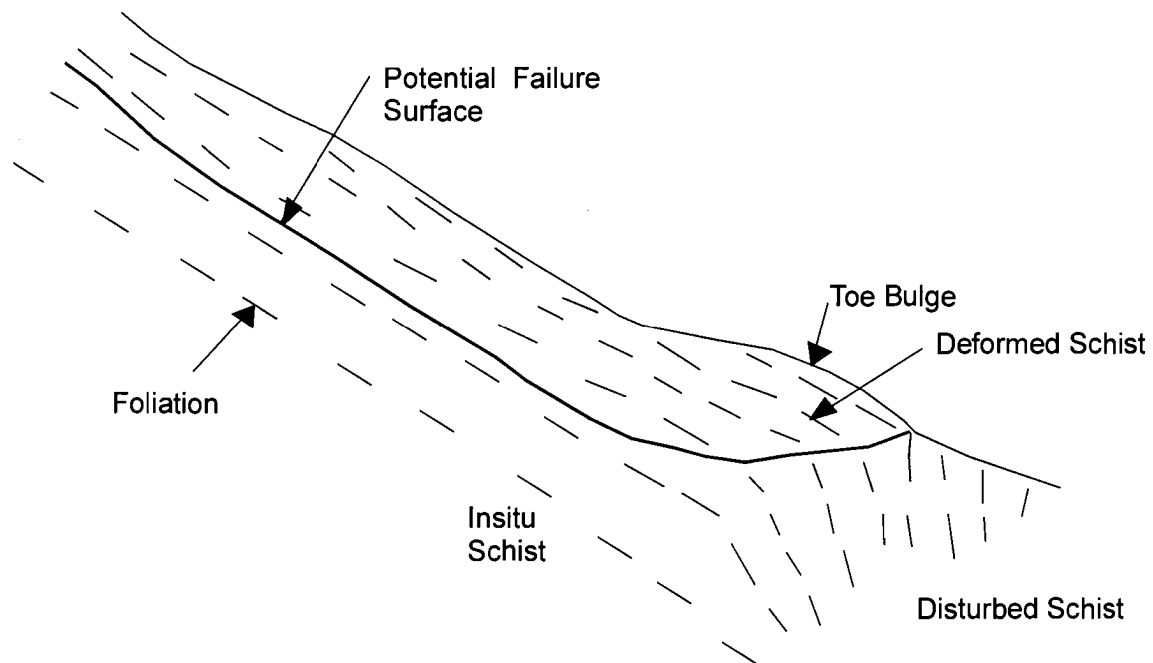


Fig 2.6 a Development of a toe buckling fold due to initial movement downslope to release stress.

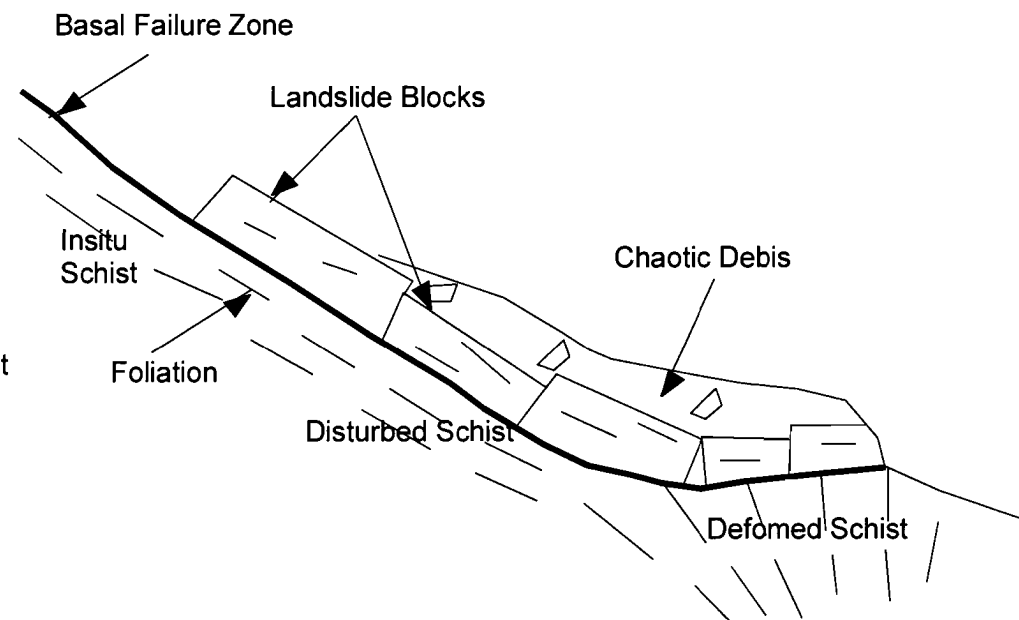


Fig 2.6b Actual failure of the slope, showing the juxtaposition of flat lying schist over near vertical schist.

2.2.2.2 Other Defects

Many defects in the schist are a result of the tectonic activity that the rock mass has undergone. Faults are crushed zones up to several 10's m thick (locally to 100mm), and crushed and fractured rock can persist for many meters away from the principal zone of deformation. Hanging walls, footwalls and internal gouge seams, which are distinctive of faults, are all present.

Crushed and sheared zones with smaller apparent fault offsets, and joints of similar attitude, typically form defect sets sub-parallel to the mapped major faults (MacFarlane et al 1992). Foliation shears zones of clay-rich crushed schist are usually <100mm thick.

These shear zones form sub-parallel to the foliation during metamorphism and folding, and are common in throughout the schist region. They include crushed or sheared zones, resulting from flexure in response to both folding and direct shear displacements in response to faulting (MacFarlane et al 1992). Very thin gouge seams can be found as narrow seams within the crushed material. The shears are typically 50mm to 2m thick and spaced 5 to 50m apart. Locally the foliation shears splinter, warp, terminate against other defects, or die out.

Stress relief features are generally recognised by opening of defects resulting in reduction in rock mass quality. Extensional joints infill zones and conduits are all developed as a result of rock mass dilation (MacFarlane et al 1992). In the weaker mica-rich greyschists significant flexural deformation (i.e. having undergone several phases of folding), due to stress relief, is recognised.

2.2.2.3 Shear Strength Parameters

The field strength of typical defects is derived from two components: the residual friction angle and waviness effects (MacFarlane et al 1992). Residual strength is dependent on the presence of clay gouge seams, which have typical values of $\phi = 8$ to 14° for samples of highly plastic clay.

Waviness is almost always present and may contribute between 10 and 50% of the field strength. The reason for this is that most defect walls have asperities, which interlock together and give some frictional strength. When these defect walls move past one another they initially produce a friction effect, but once this strength is exceeded they shear against each other and get planed off. Joint strength is usually measured in the field with a Schmidt Hammer.

2.2.3 *Landslide Morphology*

2.2.3.1 Asymmetric Valley Profile

Landsliding in schist terrain is most likely to form on slopes where glacial erosion or fluvial incision has carved out the valley, leaving it with very steep sides. Once the glacier retreats this leaves the sides unsupported, and landsliding develops on the slopes where foliation dips into the valley. Where this happens the slopes are gentler ($20 - 25^\circ$). On the other side of the valley, however, the slopes are steeper and more uniform as the foliation dips into the rock (Fig 2.7). This produces an asymmetric profile for the valley.

An example that illustrates this asymmetry occurs at the Paerau Diversion Works Scheme in Maniototo, a combined hydroelectric and irrigation scheme, where the canals were built in the Otago Schist.

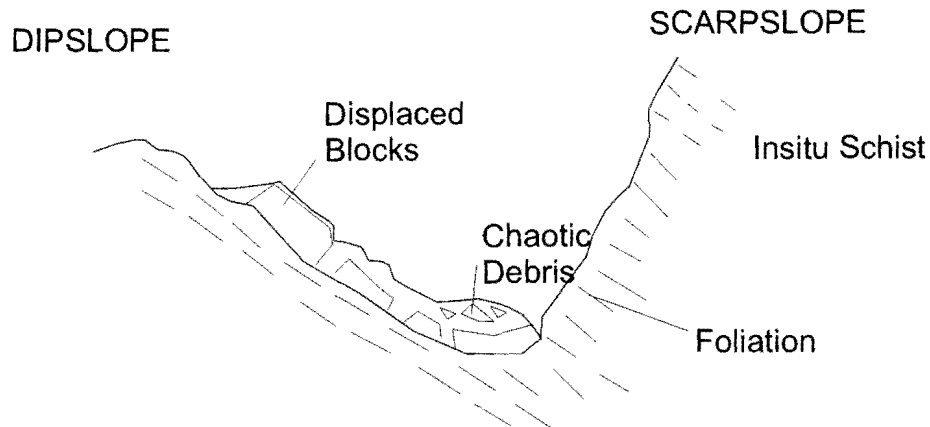


Fig 2.7 Cross-section of an asymmetric profile where sliding has developed.

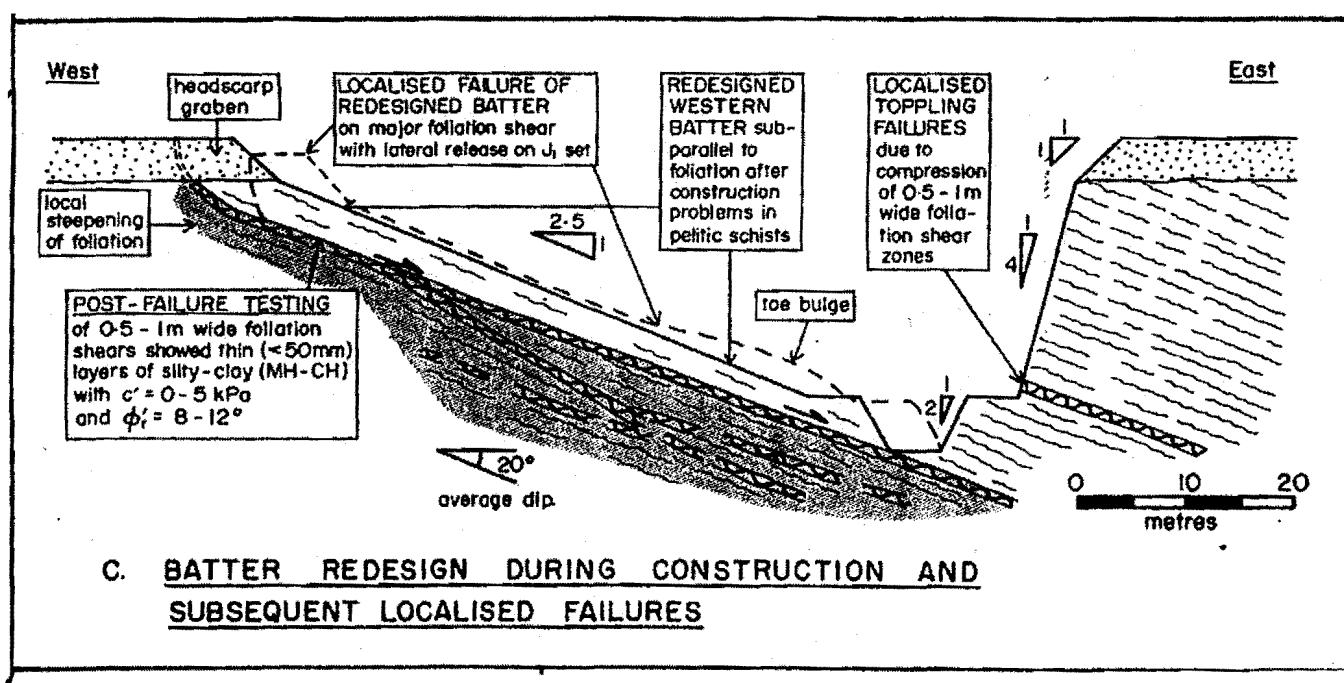


Fig 2.8 Cross-section of the failed batter at Maniototo, showing the asymmetrical profile that was adopted for this project. From Moody (1985).

When the canals were built, the western batter failed along a major foliation shear zone with lateral release on a joint set, after the batter had been redesigned with a gradient close to the dip of the schist foliation (Fig2.8). The batter failed because the foliation dipped into the canal, and the gradient of the batter allowed the foliation shear to daylight in the slope. As the foliation dipped into the slope face on the eastern batter, there was no danger of it failing.

2.2.3.2 Volumes and Dimensions

The volumes of the landslides in general can vary greatly. Planar failures can range from a few cubic metres to large-scale failures that involve entire mountainsides. Looking specifically at landslides in Central Otago, the smallest is Miners Rockfall in the Cromwell Gorge ($0.75 \times 10^6 \text{m}^3$) and the largest is the Coronet Peak Landslide in the Wakatipu Basin ($1.5 \times 10^9 \text{m}^3$). They each have areas of 5ha to 1800ha respectively. Typically landslides can extend for several kilometres or more along the valley side, and usually from ridge crest to valley floor. Most are deep-seated failures, so maximum thicknesses (vertical depths to slide base) would range from 30-150m.

2.2.3.3 Surface Features

A usual surface feature of these slides is 'hummocky' or 'ripple' terrain. This is the characteristic surface expression of deep-seated failure in the schist, which occurs over a period of thousands to several thousands of years. Also called sagging metamorphic terrain, it is said that this terrain is characteristic of ranges that have recently been uplifted or experienced glacial erosion.

Sagging metamorphic terrain refers to mass movements that form a stepped and rippled landscape, underlain by deep-seated rockmass creep along structural defects present in the rockmass.

Fretting of the bedrock exposed in the headscarps can occur, resulting in the slopes of the landslides being littered with schist blocks. Surface water, in the form of streams is another characteristic of landslides. Tarns and seepages are also common.

2.2.3.4 Subsurface Features

Usually streams on the surface of the landslide are formed from seepage of confined groundwater or water trapped beneath the slide when it moved so groundwater is an important subsurface feature. This can be as perched or confined and semi-confined aquifers. Confined and semi-confined aquifers are the most important, as they can lower the overall stability of the slide (eg Jacksons Creek Slide, New Zealand).

Foliation shear zones are another important subsurface feature, and more often than not form either the basal failure zone or internal shear zones. The Basal Failure Zone is another subsurface feature that is important, and sometimes there are one or several failure zones above the basal one (eg Downie Slide). Basal Failure Zones occur with any mechanism of sliding and usually consist of thin seams of clay gouge.

If the failure mechanism is toe buckling, and then steepened to overturned schist will be observed in the subsurface (eg Nine Mile Slide, New Zealand).

2.3 Landslides in Central Otago

2.3.1 Geological Setting

Bedrock for the Central Otago region is the lithologically variable Otago Schist. Having undergone several phases of folding and faulting the schist has many defects (such as joints, foliation shears, crushed zones, and fault zones). The foliation is well developed and sub-parallel to the schistosity. Failures in the schist are common, mainly planar, and occur on slopes where the foliation dips into the valley at about 25°.

The main deposits of the Tertiary are the Manuherikia Group, and they consist of poorly cemented layers of sand, mud and lignite. There are also the Bobs Cove Beds, which are marine sandstones, limestones and conglomerates that were deposited during a brief incursion of the sea. The deposits of the Tertiary are poorly preserved, with remnants occurring in several places only, mainly as inliers along faults.

As there is so little Tertiary around in the Central Otago region, landslides in this material are rare. The Resta Road Slide is the only failure known to the author.

Reactivation in the regolith material of the landslides is quite common. This material consists of chaotic, predominantly matrix supported (silt and sand) schist debris. Occasional schist blocks are visible in the slide debris.

There are several reactivations on Coronet Peak Landslide, across from Arthurs Point, and near or on the ski field. On Arthurs Point Landslide there are several zones of more active movement. In the Wakatipu Basin there are only failures in the bedrock, and reactivations in the debris. No Tertiary sediments are present in the basin for failures to occur in.

2.3.2 Major Schist Slides

Central Otago covers a large area, and within this many landslides can be found.

In general most landslides are foliation-controlled translational features (eg Coronet Peak). They are large, old, and usually creeping so they have the characteristic 'hummocky terrain'. Wedge failures also occur, where a combination of defects intersects with one another (eg Arthurs Point). As these landslides are old they have subdued morphology associated with cryergic smoothing.

Toe buckling is a common failure mechanism, where gravitational relaxing of the schist occurs after initial oversteepening through glaciation and is then triggering by further glacial or fluvial processes. Initially a fold forms then through gravity the mass shears through the zone of maximum deformation. Further movement downslope places flat-lying schist over 'sub-vertical' schist. A zone of steepened to overturned schist outcrops mostly on the left bank of the Cromwell Gorge, and is present in the toe areas from Nine Mile upstream through to Dunlays Slide (Fig 2.9). Table 2.1 lists a selection of the major schist slides of Central Otago.

2.4 Lake Dunstan Landslide Project

2.4.1 Nature and Extent of the Landslides

Many of the slides have relief of 300 to 500m, slope lengths of one to several kilometres and most are developed on the left (NE) side of the valley (partially a foliation dipslope), but slides also occur on scarp slopes too (eg Clyde Slide and Cairnmuir Slide). Areas range from 5ha (Miners Rockfall) to 900ha (Nine Mile Downstream), which have volumes of $0.75 \times 10^6 \text{m}^3$ and $>1000 \times 10^6 \text{m}^3$ respectively.

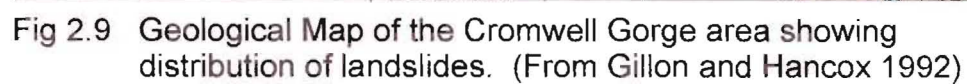


Fig 2.9 Geological Map of the Cromwell Gorge area showing distribution of landslides. (From Gillon and Hancox 1992)

TABLE 2.1 : Summary of Physical Characteristics of Major Landslides in the Wakatipu Area

Slide Name	Area (ha)	Volume 1*10 ⁶ m ³	Max. Thickness (m)	Average Slope Angle (o)	Comments on Slope Movement
Arthurs Point	80	40	70	40	SS; wg on jts
Brewery Creek	200	175	140	26	SS; cz ctrl
Byford Creek	7	2.8	50		
Cairnmuir	100	10	83	25	
Clyde	120	50	70	22-28	DS; cz (f), stp
Cornish Point	13	4	20	30	DS; wg on (f)
Coronet Peak	1800	2700	200	25-30	DS; (f)
Cromwell	7.5	3	35	31	DS; (f), jt ctrl
Devils Creek	250	375	150	24	DS; (f)
Dunlays	74	55	120	23-32	DS; low angle
Flying Fox	14	5	60	27	DS; low angle
Gibbston	600	120	50		DS; jts, (f) wg on fault
Hintons	22	8	70	34	DS; (cz)
Jacksons Creek	23	5	46	28-30	SS; (cz)
K9	950	500	150	20-30	DS; (f)
Miners (Rockfall)	5	0.75	30	40	RF/top; (f), jt ctrl
Muddy Creek	300	60	50		DS; jts, (f) wg on fault
Nine Mile	900	>1000	200	16-27	DS; (f) + czs
No 5 Creek	126	60	100	19-35	DS; (f), jts + cz
Queenstown Hill	200	30	20		DS;
Ripponvale	150	20	20		
Roaring Meg	740	370	150	20-30	DS; (f)
Two Bridges	21	7	80	30	DS; (f)

NOTE:

- 1) Ranges given for slope angles generally indicate flatter upper slopes, and steeper lower slopes
- 2) Abbreviations used are as follows: (DS)-dip slope; (SS)-scarp slope; (f)-foliation or foliation shear; (jt)-joint; (cz)-crushed zone; (ctrl)-controlled; (stp)-stepped; (wg)-wedge failure.

Thicknesses range from 20 to 200m, while average slope angles range from 16° to 40°. Exhibiting subtle, subdued morphology most of the landslides are ancient features that lack prominent denuded scarps or obvious hummocky topography.

The slides are the result of long term deep-seated creep of schist bedrock, colluvium and solifluction deposits (Fig 2.9), particularly on dipslopes formed parallel to foliation. Extending from ridge-crest to valley floor, the toes of some of the slides are moving into the river channels in response to erosion.

Most slides in schist bedrock and colluvium are considered to be creeping translational rock and chaotic debris slides developed along foliation surfaces, foliation shears, fault crush and shear zones, and joints (Gillon and Hancox 1992). Landsliding in the Cromwell Gorge has developed gradually during the Quaternary in response to the incision of the gorge by the Clutha River and tributary streams through the uplifted Dunstan-Cairnmuir mountain block (Beetham et al 1992).

2.4.2 Investigation and Remediation

Detailed surface geological mapping was carried out over a period of 17 years, along with sub-surface and surface geotechnical investigations to define the activity and extent of the Cromwell Gorge slides (Gillon and Hancox 1992). Therefore the development of the Clyde Power Project has involved a very in-depth investigation of the slides.

From 1983 to 1986 geological investigations were carried out in the Kowarau Valley and on the new highway route through Cromwell Gorge, which showed that 'dormant' schist slides in the Cromwell Gorge could be creeping. Installation of surface pillars and a programme of monitoring and drilling investigations confirmed creep movements (an average of 12mm/yr) for several of the slides during 1985-1987.

In 1986 a complex high-pressure groundwater system was discovered in No 5 Creek slide due to drilling for the new highway. This was significant, as previous studies (mainly in relation to Clyde, Cromwell, and other left bank slides) had indicated that the slides had simple low gradient groundwater systems. The discovery of this groundwater system meant that lake-filling effects would have a greater influence on the stability of the landslides than previously thought.

This led to a very extensive drilling programme to define the groundwater systems of the other slides, with similar conditions being found in Jackson Creek, Nine Mile and Dunlays slides (1988-1989).

As the need for remedial works increased, more detailed geologic and geomorphic mapping of the slides was needed. The period between 1989-1990 involved an intensive phase of investigation in order to develop a realistic slide model. Because the model was developed from extensive information from many landslides, it can be correlated to other slides in the region that have similar properties.

Of all the remedial works required drainage measures are widely recognised as the single most effective and economic stabilisation measure for large rockslides (Gillon and Hancox, 1992). Although the use of sub-surface drainage wherever possible is advised, it can be supplemented by other measures such as gravity drainage for slides with extensive groundwater, and buttressing of the toe area. The general locations and extent of the drainage systems for the Cromwell Gorge landslides are shown in Fig 2.10 (Gillon and Hancox, 1992).

At the Brewery Creek Slide low-level drainage has been used on the downstream segment to reduce the water level in a semi-confined aquifer within the slide. The aquifer has an exceptionally flat gradient up to 600m back from the river (Fig 2.11). Because there was no significant drainable head in the slide, gravity drainage was not feasible. A zoned earthworks blanket and grout curtain were also used to act as barriers to flow from the reservoir to the drainage works, and all the remedial works have been designed to achieve drawdown to the post lake-filling phreatic surface (Fig 2.11).

Three parameters of the slides have been monitored. Surface deformation is monitored by 248 survey points, initially installed to determine rates of movement. Inclined meters, initially used to help locate active failure surfaces, monitor subsurface deformation. Piezometers were initially used to identify aquifers and measure groundwater levels, but are now used to monitor groundwater pressure. A summary of all investigations and monitoring of all slides is shown in Table 2.2.

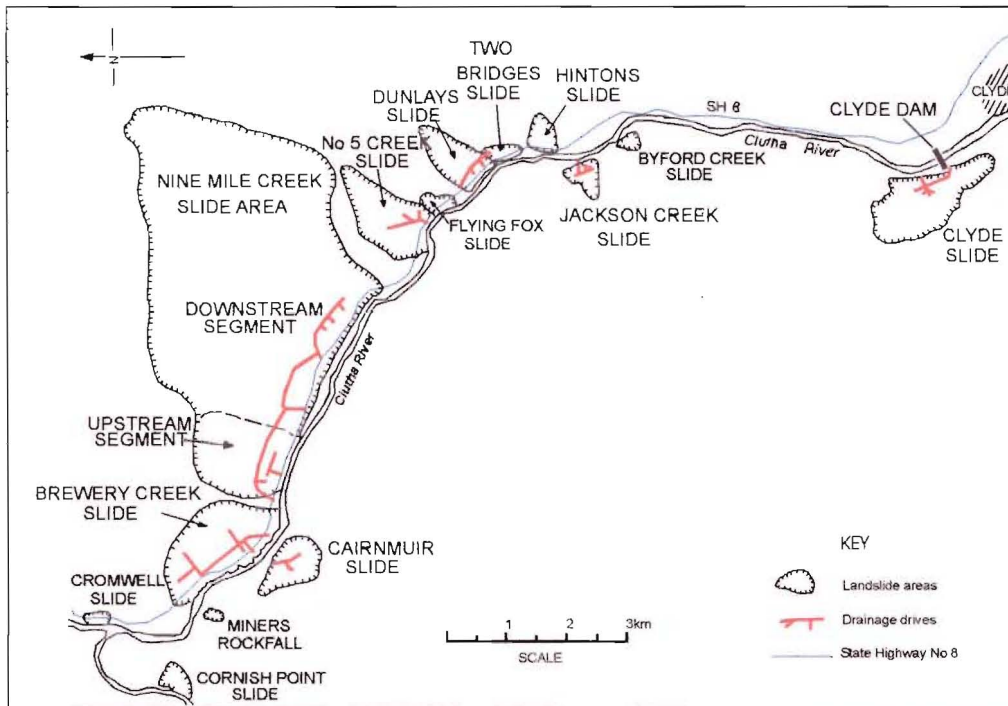


Figure 2.10 Map of the Cromwell Gorge landslide areas showing locations and extent of drainage drives.
(Figure drafted from figure 6, Gillon and Hancox 1992)

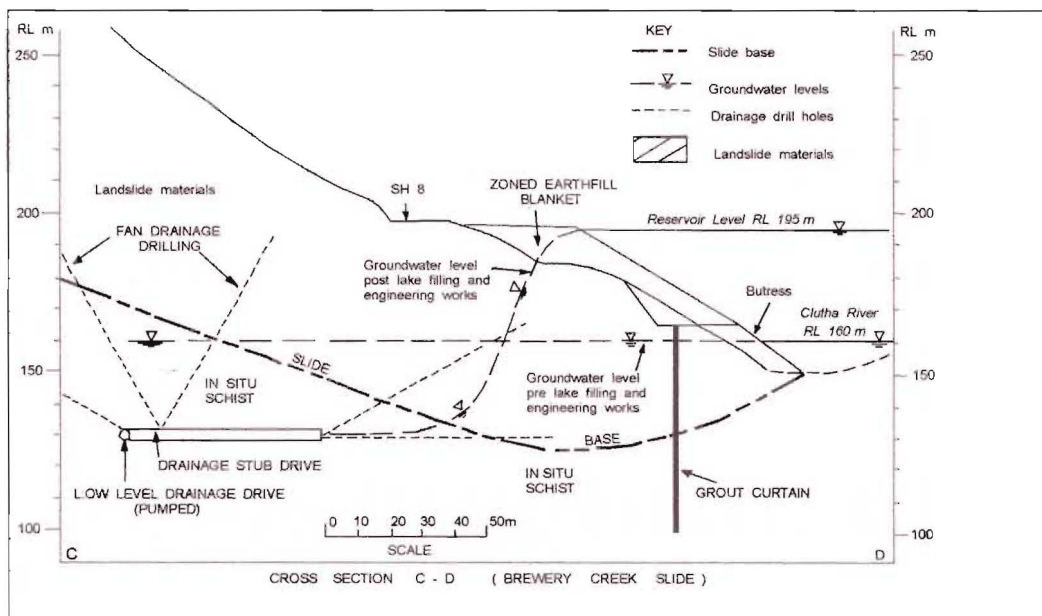


Figure 2.11 Stabilisation by low level pumped drainage, grouting, and blanketing works at Brewery Creek Slide
(Figure drafted from Figure 8, Gillon and Hancox 1992)

TABLE 2.2

Summary of Subsurface Investigations and Monitoring of the Cromwell Gorge Landslides

Landslide	Drillholes		Shafts		Test pits & Trenches No.	Seismic Lines		No. inclino -meters	No. piezo- meters	No. survey points	Level lines (m)	Drives	
	No.	Total length (m)	No.	Total length (m)		No.	Total length (m)					No.	Total length (m)
Clyde	71	5168	1	23	6	7	3959	24	53	21	925	6	1240
Byford Creek	1	74	4	45		1	470	1	1	1			
Jacksons Creek	30	1829	6	129	18	10	4719	19	66	30	940	5	812
No 1 Creek	7	546	1	13	12								
Hintons	11	696				2	830	5	11	5			
Two Bridges	15	911			19	10	3419	6	17	5			
Dunlays	23	1786			16	6	2928	8	29	7			
Flying Fox	10	513			1			4	11	5		3	1121
No 5 Creek	30	2646				6	1239	13	41	8	1207		
Nine Mile DS	96	14091	10	199	42	21	19801	19	218	38	1900	4	710
Nine Mile US	71	8210	1	28	37	10	8680	19	195	23		7	4850
Cairnmuir	25	2442	1	28	12	5	3790	5	120	15		9	3283
Brewery Creek	103	10208			36	18	11125	22	142	37	6120	2	680
Cromwell	21	1083					617	8	14	17		12	3622
Miners (Rockfall)	1	44				1	133	1	1	2		1	74
Cornish Point	2	132						2	3	5			
Bannockburn	6	299	1	18	8			1	9	5			
Pipeclay Gully	9	395			5			2	12	11			
Ripponvale	5	362				1	498	5	6	10			
Northburn	10	435			52			4	15				
First time slide areas	13	570						10	28	3			
Total	560	52440	25	483	264	51	62208	178	992	248	11092	49	16392

(Table 2.3 from Gillon et al 1992)

2.5 Large Landslides of the Kawarau Valley and Wakatipu Basin

2.5.1 *Kawarau Valley*

2.5.1.1 K9 and Roaring Meg Landslide Complex

The K9 Landslide is located in the lower Kawarau Valley, Central Otago (Fig 2.12), and it covers an area of about 950 hectares on the southern flank of the Pisa Range (Bell 1987). Extending 4km along the northern side of the Kawarau Gorge and 800m above the present valley floor, it has an estimated volume of $5 \times 10^8 \text{ m}^3$. It forms only part of an 1800(+) hectare mass movement complex, extending for 9km along the Kawarau–Roaring Meg alignment (Fig 2.12) (Bell 1987).

The Roaring Meg segment extends along the northern side of Roaring Meg to where it meets with Evan Roberts Creek (Fig 2.12). Its western boundary is the north-trending part of Roaring Meg, which is aligned approximately parallel to the Roaring Meg Fault Zone (Fig 2.12). The eastern limit is what was thought to be an insitu schist ridge separating the K9 and Roaring Meg Landslides (Fig 2.12), but evidence shows major slope movements within the ridge have led to partial failure. In the lower part of the ridge there is obvious block rotation, which provides evidence that toe buckling, and partial over-riding by massive blocks displaced from upslope, has occurred.

Two major topographic depressions have been identified, varying in width up to 100m. The margins have been subdued due to infilling by freeze-thaw-derived schist colluvium. These depressions are interpreted as graben features that developed when toe buckling occurred and the rock mass was displaced by approximately 100m.

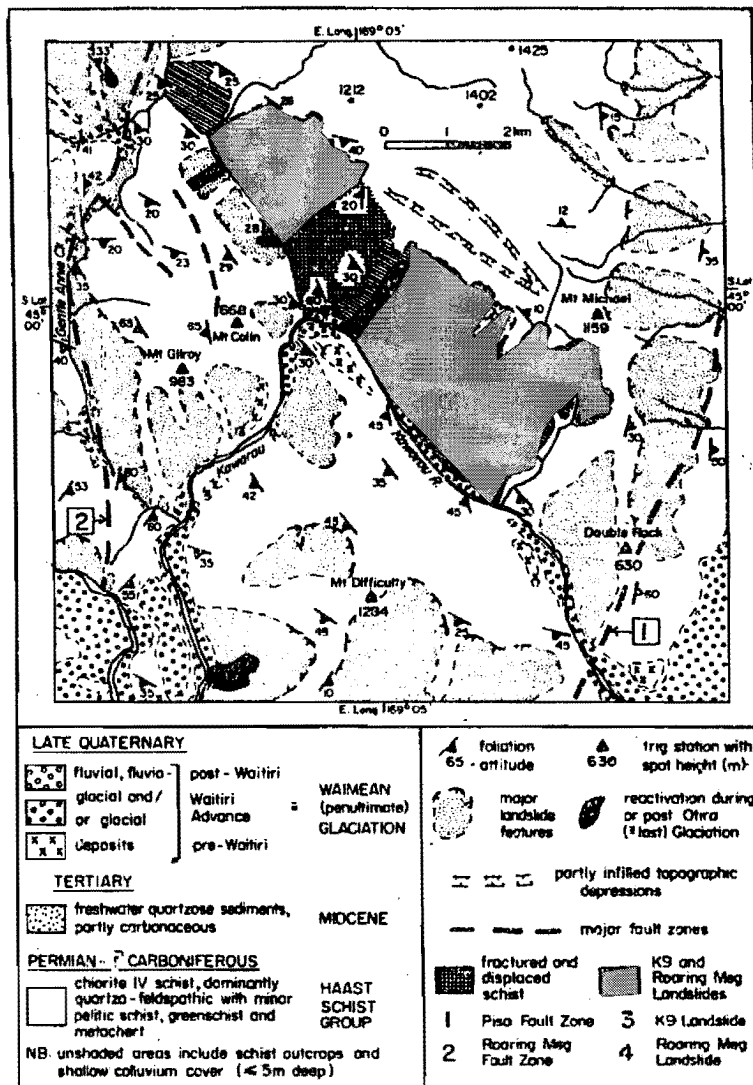


Fig 2.12 Map of the Kawarau Valley showing the location of the K9 and Roaring Meg slides. From Bell (1987).

This in turn implies foliation-controlled shear displacements beneath the upper slope segment, but the nature of the movement beneath the lower part is unknown as no subsurface investigations were undertaken (Bell 1987).

It could also be argued that the topographic depressions represent gravitational spreading features (sackung), which formed during the relatively rapid uplift of the Pisa Range in the Quaternary, due to reverse movements on the Pisa Fault Zone. If gravitational spreading of the Pisa Range has occurred during Quaternary tectonism, then the observed toe buckling (which is clearly a precursor to large scale slope failure) can be attributed to bulging at the lower slope consequential on progressive creep along foliation surfaces beneath the upper slope segment (Bell 1987). Some form of toe support removal, like glacial over-steepening or river incision, would achieve triggering.

Suggested Failure Model (From Bell 1987)

- 1) Relatively rapid uplift of the Pisa Range occurred during the Quaternary as a consequence of episodic movements along the Pisa Fault Zone, accompanied by deep valley incision on structurally - controlled bedrock courses under fluctuating climatic conditions and base level controls.
- 2) Gravitational spreading along the southern (and probably also the northern) flank of the Pisa Range then followed, with toe bulging and headscarp graben development over a vertical interval of some 1000m and for a distance of more than 9km along the valley sides.

- 3) Major foliation-controlled rock slope failures (giving rise to the K9 and Roaring Meg Landslides) occurred during the Late Quaternary (and prior to the Waimean Glacial Stage), with triggering by one or more of fluvial incision in the toe areas, ice over-deepening of the Kawarau Valley floor, and earthquake shaking accompanying ongoing tectonism.
- 4) Significant periglacial modification of the landslide and valley slopes during the Waimean and Otiran Glacial Stages, with the initiation of secondary (and tertiary) failures within the existing slide debris.

2.5.1.2 Gibbston Slide Complex

The Gibbston Slide is a large mass movement area on the northern face of the Gibbston Basin in the Kawarau Valley, Central Otago (Johnson 1986). With a surface area of about 6km^2 and an estimated volume of $1 \times 10^8 \text{m}^3$ it is the second largest mass movement in the valley. The Muddy Creek Slide is separated from the Gibbston Slide by a narrow ridge of insitu schist (Fig 2.13), and has an approximate surface area of 3km^2 and an estimated volume of $1 \times 10^7 \text{m}^3$.

In the Gibbston Slide, failure has occurred along dislocation surfaces formed by defects oriented unfavourably (such as foliation shears, mica-rich horizons, foliation and joints). In the downstream slide segment the failure model is one of dipslope failure on either foliation shears or weak mica-rich horizons (Johnson 1986). Upstream there is a transition from pure dipslip to a complex dipslip/wedge failure, formed by the intersection of the foliation with prominent joint sets. The Gibbston Fault (a reverse fault) has also had some influence, forming a dipslope/wedge failure between the fault and schistosity, with sliding along the fault crush zone (Johnson 1986).

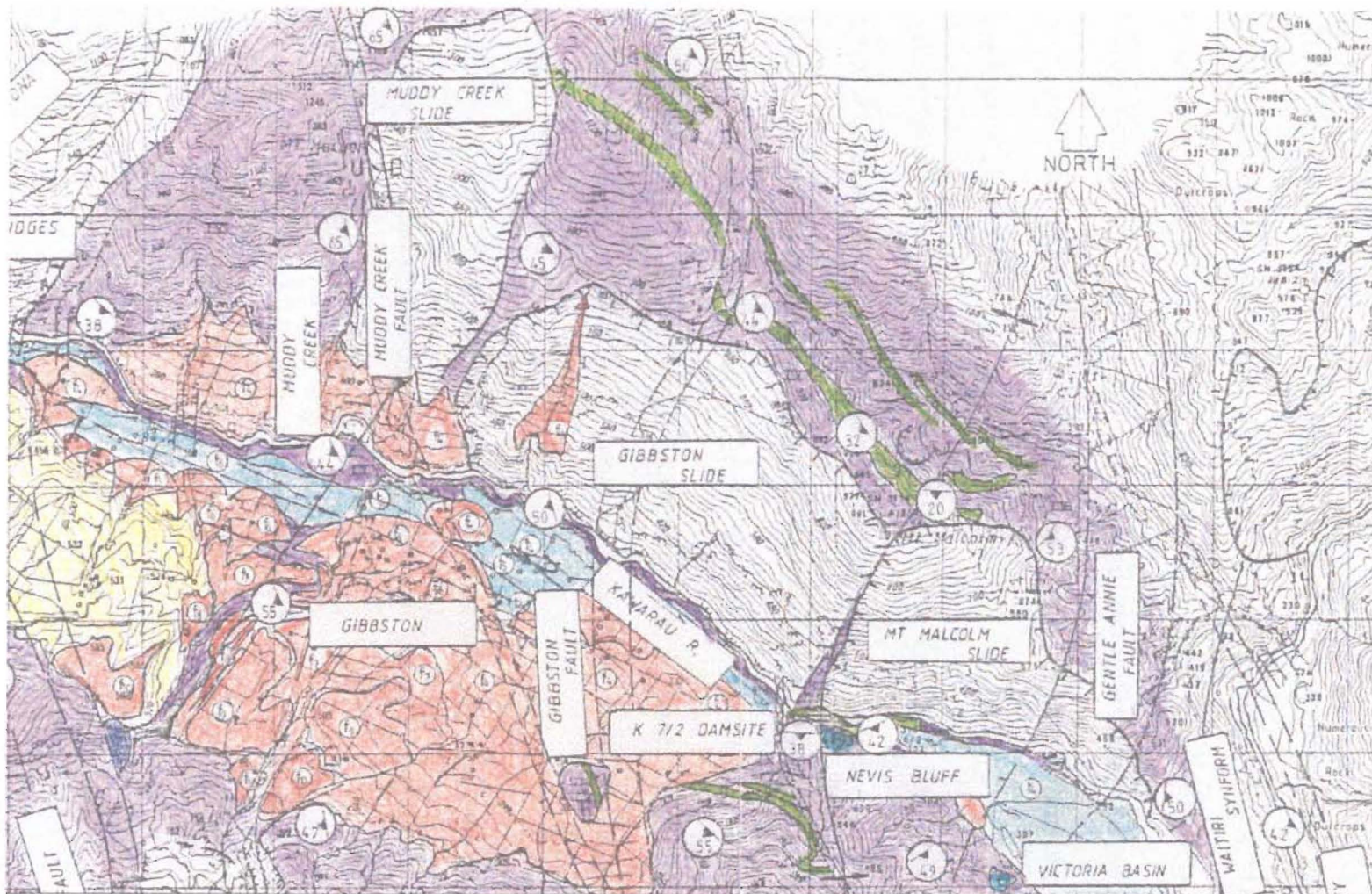


Fig 2.13 Map of the Gibbston Valley showing the Gibbston and Muddy Creek slides.
From Johnson (1986).

Natural steepening of schistosity downslope (due to the influence of the Mt Rosa Antiform) results in the formation of a convex failure profile. This results in oversteepening of schist blocks with progressive downslope movement, and the tendency for juxtaposed blocks to under or over ride one another (Fig 2.14).

The failure has formed as a result of the removal of basal support from the slope, related to the retreat of glacial ice in the Late Quaternary. Progressive failure of the slope occurred throughout the Late Pleistocene and Recent in response to changes in river base level.

The age of the slide was estimated by projecting the surface of an alluvial fan remnant that was deposited on the slide to an aggradation base level in the axis of the Kowarau Valley, and correlating this with a projection of the aggradation fan surfaces f4 to f10 preserved in Camp Creek. The fan on the Gibbston Slide projects to the f8 fan surface of Camp Creek (Fig 2.15; Johnson 1986).

The fan surface was tentatively dated at 36 000yrs BP, so the slide is estimated to be older than this.

Luggate ice (70 000 BP) is considered to have excavated the main Gibbston Basin, so it is concluded that initiation of the Gibbston Slide is probably related to the retreat Luggate ice (70 000 BP).

In the Muddy Creek Slide failure occurred along the same kind of zones as in the Gibbston Slide, with the same controls (foliation shears, schistosity and the Muddy Creek Fault).

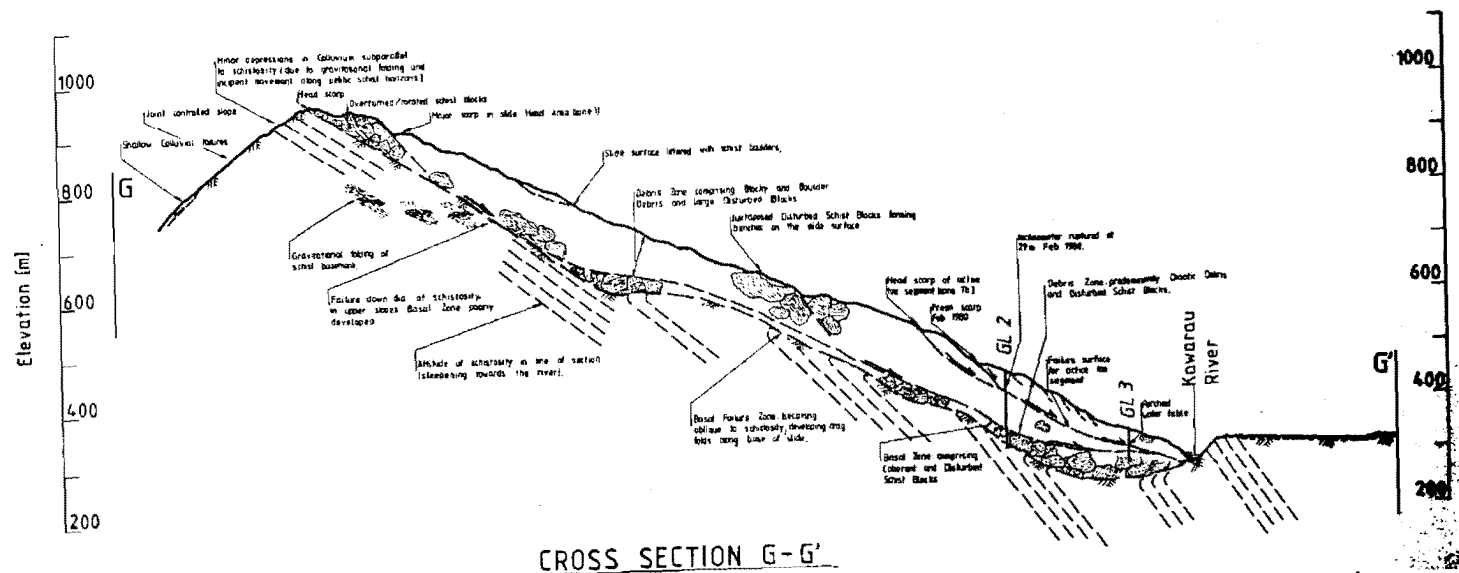


Fig 2.14 Cross-section of Gibbston Slide showing juxtatposed blocks.
From Johnson (1986).

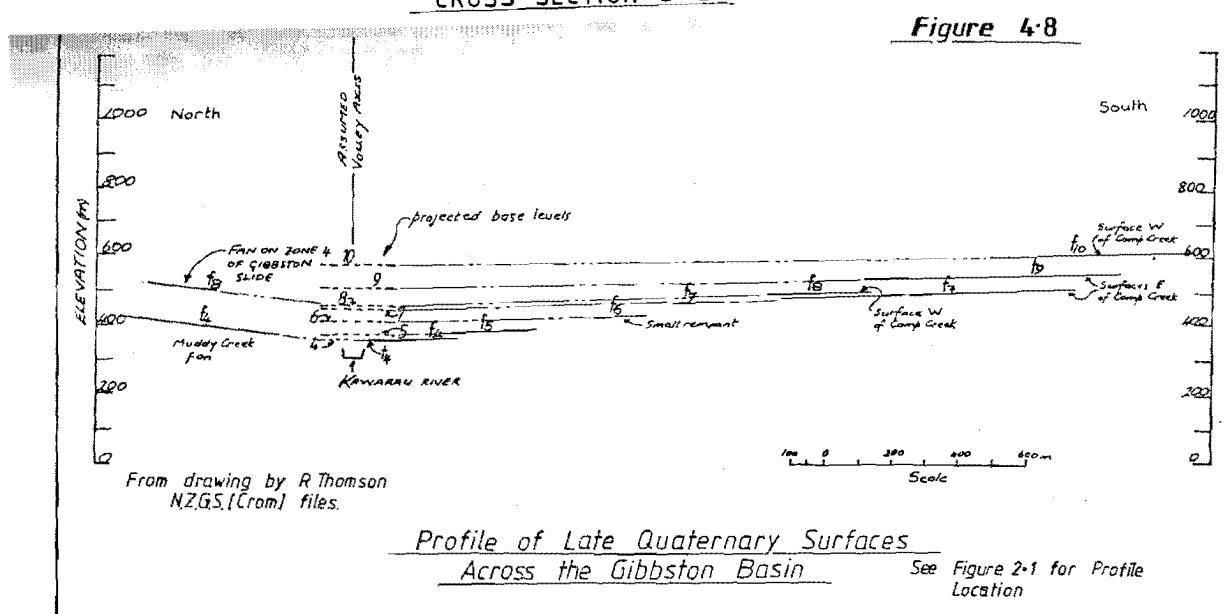


Fig 2.15 Profile of Late Quaternary surfaces across the Gibbston Basin.
From Johnson (1986).

2.5.2 *Wakatipu Basin*

2.5.2.1 Coronet Peak

The Coronet Peak Landslide is a creeping translational slide formed in the schist bedrock of the Central Otago region (Fig 2.16). It has an estimated volume of $1 \times 10^9 \text{ m}^3$ and forms the northwestern boundary of the Wakatipu Basin.

This slide is a foliation-controlled planar failure occurring either along foliation or foliation shear zones.

It is likely that the triggering mechanism is the excavation of the valley by glacial erosion and then removal of support when the glacier retreated. No over steepened schist has been observed, so toe buckling is inferred not to have happened.

2.5.2.2 Arthurs Point

Arthurs Point Landslide is situated on the northeast-facing slope of Bowen Peak, Queenstown (Fig 2.17) and has an estimated volume of $4 \times 10^7 \text{ m}^3$. It is thought that the landslide failed as a wedge, most probably on the intersection of two sets of joints.

The topography of the slide is extremely steep (40°), with hummocky terrain not well-developed, though it is present.

Triggering of the failure was likely to have been caused by incision of the Shotover River, which continually erodes the toe of the slide, resulting in the continual creep of the landslide downslope.

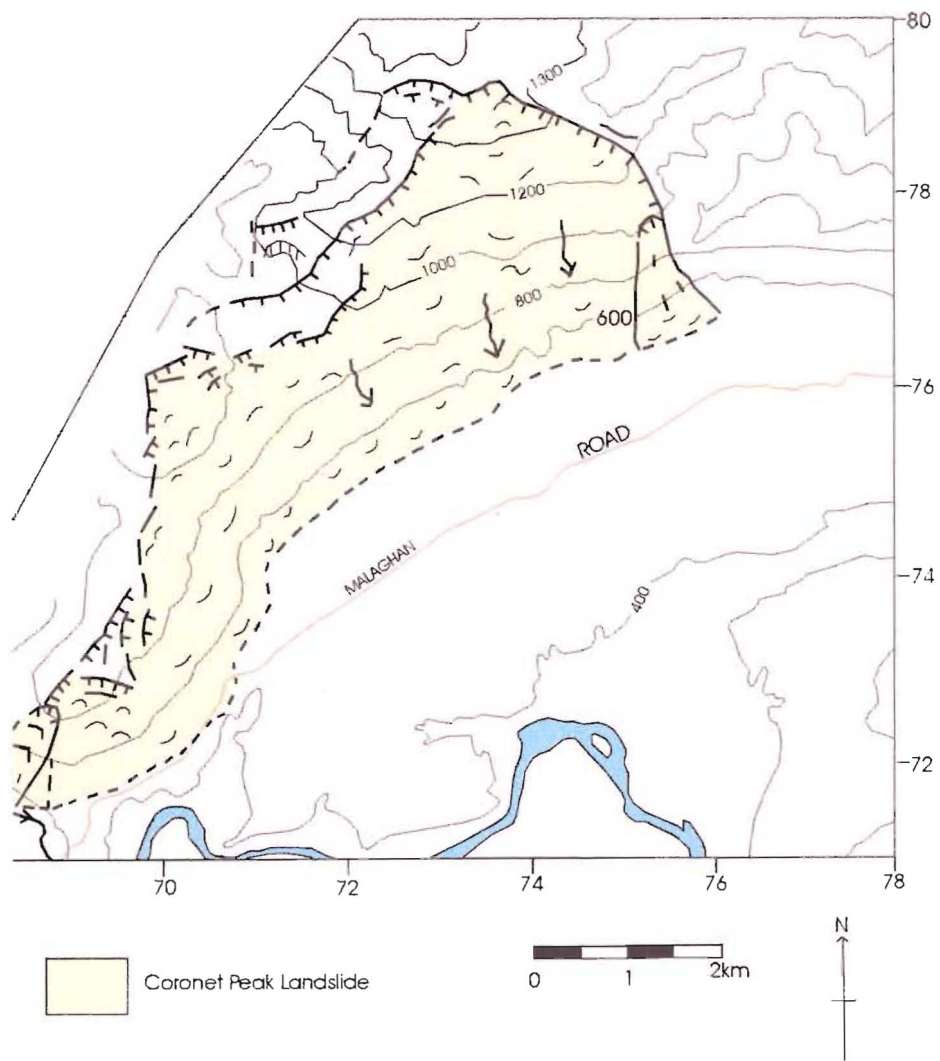


Fig 2.16 Coronet Peak Landslide

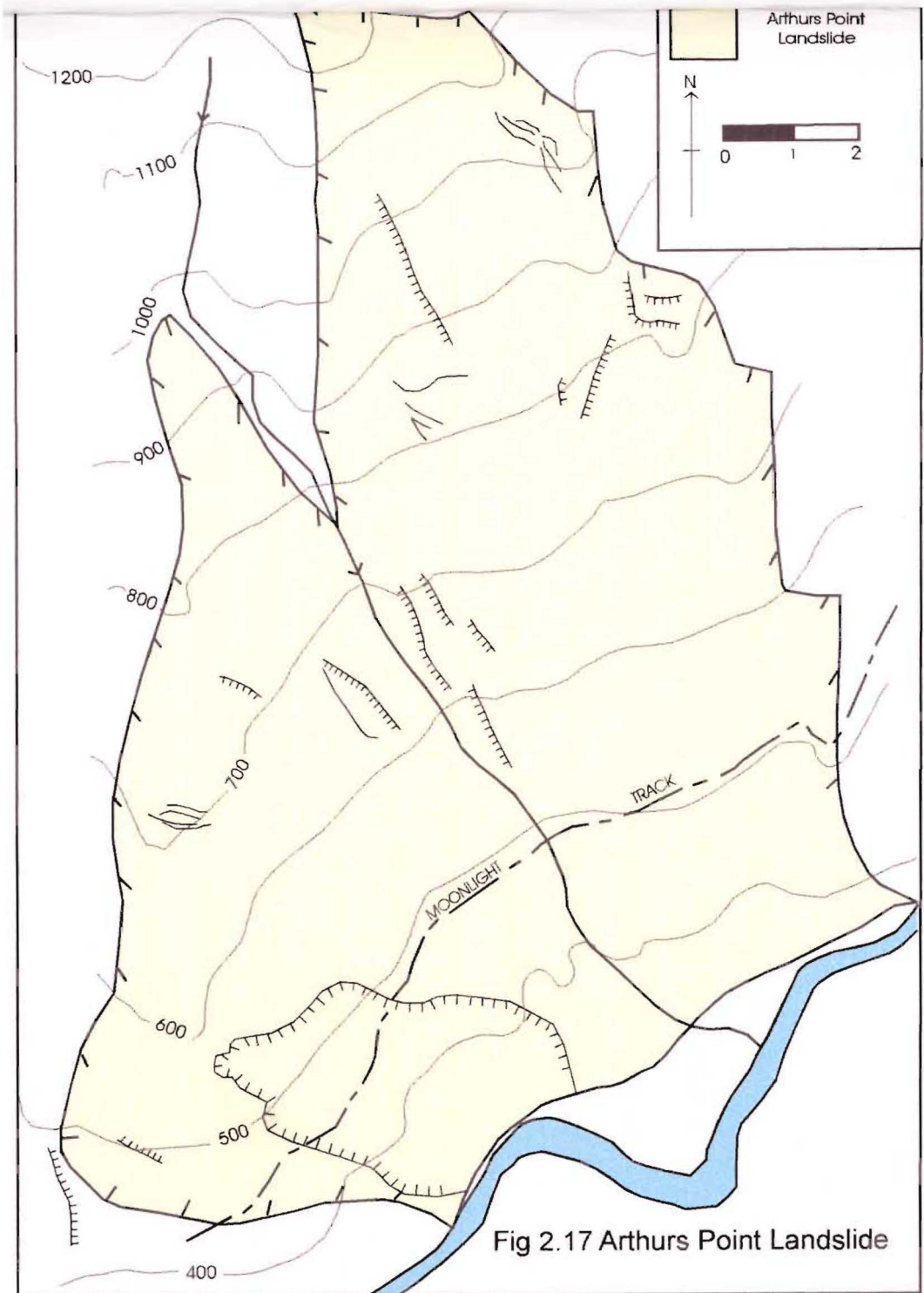


Fig 2.17 Arthurs Point Landslide

2.5.2.3 Queenstown Hill

Queenstown Hill Landslide is a smaller translational slide, located on the southeast slope of Queenstown Hill. Extending from 20m below the summit of Queenstown Hill to about 460m above sea-level, it has an area of 2km² and an estimated volume of 2.4*10⁸m³. The well defined but degraded headscarp is joint-controlled, along with several smaller scarps upslope that represent the retrogressing of the headzone. Immediately downslope a graben runs parallel to the headscarp. A veneer of till covers the headzone, and so limits outcropping of the schist. A prominent joint-controlled lateral scarp is to the west, with several grabens and open cracks present in the upper slopes. The toe of the slope is difficult to detect, and is inferred to occur at 460m above sea level.

The upper half of the slide mass contains many well-developed scarps, and others that appear poorly defined. Large chasms (void volumes of tens of cubic metres) occur in the central slide mass, along with wide, open shear cracks. Most of the slide shows hummocky topography and chaotic relief, indicating significant movement, and is well developed in the southwestern area.

As no subsurface investigation was undertaken, depth to the failure surface is inferred to be 100-150m. Failure is likely to have occurred along or parallel to schist foliation, or on foliation shears. It is inferred that failure is translational, with the toe forming a shallow compressional bulge. The shear cracks are an indication of this compression.

Queenstown Hill Landslide is a retrogressive failure, and three phases of movement have been identified. They are as follows: (From Stossell 1999)

- 1) Translational sliding in the southeastern area of the slide, as result of increased glacial undercutting at about 460m above sea level along weak pelitic schist horizons or foliation shear zones. The eastern lateral release structure would have added to instability. Several retrogressive scarps have formed upslope from this shallow failure.
- 2) Movement downslope towards the southwest to a depth of 100-150m has occurred in the northeast area as a result of removal of support in the southeast (phase 1). A series of small and very shallow translational slides formed along with compression and bulging of the toe areas. Hurrmocky topography formed during this phase. The failed area continued to retrogress upslope, with continued enlargement of the failure surface at the same time.
- 3) Extension then took place in the northwestern area to a depth of 100-150m as a result of the removal of support directly downslope, with movement towards the southeast. The insitu schist block immediately downslope below the toe area has formed a buttress, restricting further gravitational creep movement. A compression zone has resulted in the mid-slope area, containing large open chasms and shear features. This phase would have formed the retrogressive headscarp and graben.

Ice over-rode the summit of Queenstown Hill during the Kumara 2 advance, depositing the Camp Hill Formation. Although the exact age of the Queenstown Hill Landslide is unknown, it is estimated to have occurred soon after retreat of the Last Glaciation ice (approximately 18-14 000yrs ago).

2.6 Engineering Geological and Geotechnical Considerations

2.6.1 *Movement History*

It is thought that the landslides have formed as a result of ongoing tectonic (uplift) and erosional processes associated with the formation of the Cromwell Gorge over the last 1 to 2 million years.

None of the landslides show evidence of having undergone large-scale rapid movements. All available geologic and geomorphic evidence suggests that the slides have always been slow moving, creeping slope movements (rather than catastrophically rapid). Topographic profiles and the general lack of slide debris on the low terraces in the gorge indicate they have undergone little overall movement or modification over the last 500 000 years.

Landslides are currently either dormant, or creeping at only instrumentally detectable rates (less than 20mm/yr). For currently active slides geological evidence (lack of scarp development and large-scale movements on to terraces and large-scale movements on to terraces, or into the river) suggests this rate has not been uniform through time, and there are likely to have been periods of creep movement interspersed with periods of inactivity (Gillon and Hancox 1992).

Table 2.3 gives the names for the different zones, and what they contain. Slide movement is inferred from the presence of "chaotic debris" and "displaced schist", bounded at depth by a "basal failure zone". "Disturbed" and "deformed schist" are categories of subslide movement, originating through rock creep and/or stress relief (MacFarlane et al 1992). Fig 2.18 illustrates the spatial relationships.

There is no evidence to suggest that any of the landslides have moved as a single mass, and most can be zoned.

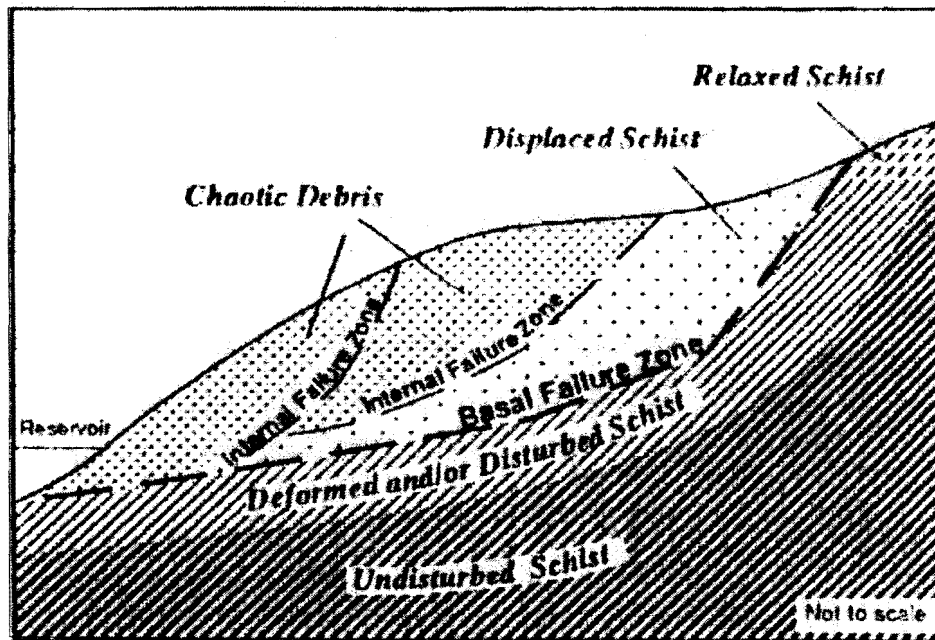


Fig 2.18 Schematic representation of general distribution of slope movement material categories, Cromwell Gorge. From MacFarlane et al (1992).

Table 2.3 Terminology Used in Slope Movement Interpretation, Cromwell Gorge.

<i>Term</i>	<i>Mass Description</i>	<i>Surface Characteristics</i>	<i>Type Of Movement</i>	<i>Degree Of Displacement</i>
Chaotic Debris	Gradation from large competent rock blocks to fine grained material (commonly intensely sheared and crushed schist), with gouge seams. Foliation attitudes of blocks normally highly variable. Slickensided downslope-dipping internal or basal failure zones.	Laterally impersistent breaks in slope (i.e hummocky). Well developed slide scarps where active. Blocks on surface.	Rotational, translational, or complex movement	10's to 100's of metres.
Displaced Schist	Large competent rock blocks either in contact or separated by open and/or infilled joints or by zones of sheared/crushed material. Foliation either parallel or oblique to undisturbed rock, and adjacent blocks may be slightly rotated relative to one another. Slickensided near downslope-dipping internal or basal failure zones, typically sub-parallel to foliation or pre-existing rock defects (Also termed "Rocky Debris", where disruption is greatest).	Laterally persistent breaks in slope (i.e broadly irregular). Forms outcrops locally.	Translational slide.	metres to 10's of metres.
Basal Failure Zone	Crushed zones and gouge seams with some sheared and shattered schist of variable thickness. Fabric may be sub-parallel to boudaries, contorted or totally disrupted. Gouge seams typically thin but persistent with slickensides, oriented downslope.	Outcrops rare.	Slide	mm's to 100's of metres.
Disturbed Schist	Sub-slide rock mass with partly open defects often infilled. Typically quartz-rich massive and laminated schist. (Termed "Relaxed Schist" in near surface situations).	Relaxed schist forms prominent outcrops similar to undisturbed schist.	Stress relief processes.	mm's to m's.
Deformed Schist	Sub-slide rock mass, fissile, with discrete sheared and crushed zones sub-parallel to flexurally deformed ("buckled") foliation, steepened to overturned locally. Typically mica-rich laminated schist.	Prominent outcrops with foliation dipping at moderate-high angles to undisturbed schist.	Bedrock flow, by stress relief and/or gravitational processes.	mm's to 10's of metres.
	In situ rock mass with closed defects	Forms prominent outcrops	None.	None.

Variation in the depth to slide base at several of the slides is indicative of steps across defects trending normal and parallel to slope direction, often approximately coincident with the boundaries of zones mapped at the surface (MacFarlane et al 1992). This evidence strongly indicates that different parts of the slide have moved at different times, at different depths and different amounts. The dominant rate of movement has probably always been slow creep.

2.6.2 Failure Surfaces

Small diameter core drilling has been used extensively to investigate the subsurface, but has rarely been conclusive in defining slope movements. In particular, recognition and correlation of weak seams forming basal failure surfaces has been difficult. Often these seams are not recovered in drillcore and their projection and interpolation between investigation points is poorly constrained. This is because there is a high degree of geological variability, complexity, and lack of continuity and marker horizons.

Curvilinear failure surfaces can be inferred locally within or at the base of the slide debris, especially in the toe areas, implying an element of rotation within an otherwise translational movement (MacFarlane et al 1992). This would be indicative of toe buckling as the failure mechanism.

The basal failure surface underlies the slide and consists of a thick zone of sheared material. Usually the failure surfaces exploit what defects there are present. Internal failure surfaces are also present, and are likely to be curvilinear in both upslope and cross slope directions. Often they have very complex interrelationships and can pond perched groundwater.

2.6.3 *Reactivation*

Most slides are probably creeping, either due to marginal strength or if they are undercut continually by river incision, further glacial activity or other landslides. In general there are several factors that can cause reactivation, some of them being seismic activity, groundwater conditions and inundation of the toes of the landslides (eg the filling of Lake Dunstan).

Seismically, the Alpine Fault is close enough to have an effect, though not as much as faults closer to or in the Central Otago area. Most of the major mountain ranges are bounded by faults, many of which have undergone repeated movements in the last 50 000 yrs (Late Quaternary) and are classed as active. It is inferred that these faults were the sources of many of the large earthquakes in the region in the last 50 000 yrs but none of the landslides show evidence for rapid large-scale movement. Therefore it is concluded that the landslides are relatively resistant to seismic shaking and active faulting. This is consistent with worldwide case histories, which suggest that most earthquake-triggered landslides are first time failures, rather than reactivations of pre-existing slides.

Groundwater is usually a very important feature of landslides, and has a great bearing on the stability or instability of the slide. Confined aquifers are the most unfavourable groundwater conditions because water can pond behind clay-rich layers or in cracks and push the landslide downhill. Also when water becomes trapped beneath or within the slide, high destabilising pressures can build up over a large area. Confined aquifers have been found at No 5 Creek Slide, Jacksons Creek Slide, Dunlays Slide and Nine Mile Slide (Gillon and Hancox, 1992).

During lake-filling associated with building of dams, many of the toes of the landslides can be submerged, which can cause problems in several ways. For example, in a previously dry landslide water can soften the dry clay minerals causing a loss of strength. Also toe submergence affects the frictional resistance in the toe area, because water decreases the effective weight of the rock.

Lake filling also affects groundwater systems within and beneath the landslide. Sometimes the groundwater system is simple, and the response to lake filling is predictable. Other times the groundwater system is complex, and the effect of lake filling is to increase groundwater pressures even more (eg No 5 Creek, Jackson Creek).

2.7 Synthesis

- Landslides in schist are common throughout the world as well as in New Zealand, and all have similar characteristics.
- The three main failure mechanisms in the schist bedrock are planar failure, wedge failure and complex failure. Planar failure is where planes of weakness are exploited. These planes can be foliation, schistosity, foliation shears, joints or faults, and generally they dip at 20 to 30 degrees into the valley.
- Wedge failures occur when defects intersect one another. Usually there is planar failure on the foliation or schistosity, with intersecting joints or faults making the wedge.

- Complex failures can be toe buckling, which is a precursor to large-scale failure. It is the downslope movement of material due to gravity that causes buckling at the toe. When this reaches a certain point it shears through and forms a proper failure.
- Toppling is another complex failure involving a combination of shear and rotation but this only becomes a factor when foliation dips at greater than 70 degrees.
- Most landslides form where foliation is parallel to the dipslope, and most develop on the left (NE) side of the valley forming an asymmetrical profile. Scarps, grabens and hummocky terrain are characteristic morphological features.
- Landslides in Central Otago mainly occur in the schist bedrock, and few occur in the Tertiary sediments as they are poorly preserved. Reactivations in the landslide debris are reasonably common, and there are several small slides in Quaternary deposits.
- A lot of investigative work was done on the Cromwell Gorge landslides, with a total of 178 inclinometers installed to help locate active failure surfaces within the slide. Nine hundred and ninety-two piezometers provided primary information on the nature and extent of the groundwater systems. Two hundred and forty-eight survey points on the slide surface were used to determine rates of movement.
- The most effective remedial measures for reservoir situations are drainage drilling measures and toe buttressing. Drainholes are there to drain the existing groundwater and to control the rise of groundwater during lake filling.

- In the Kawarau Valley and Wakatipu Basin, landslides form in much the same way as in Cromwell Gorge, and due to the same failure mechanisms and controls.
- The smallest Cromwell Gorge failure is Miners Rockfall with a volume of 750 000m³, while the largest is Nine Mile Creek Slide with a volume of less than 1*10⁹m³.
- It is very unlikely that any of the landslides moved as one mass. Instead they have probably moved different amounts at different times at different rates. Slow creep has been the dominant rate overall.
- Most of the landslides in the Cromwell Gorge are translational failures, but it is likely that the failure surface is curvilinear at the toe, which implies an element of rotation. Internal failure surfaces are likely to be curvilinear as well, and to have complex interrelationships with each other.
- Reactivation due to seismic shaking is very unlikely as the landslides have been there for 50 000yrs and there is no evidence of large-scale movement. An earthquake of a large enough magnitude would certainly have occurred during this time, and there would be evidence for its occurrence. It is therefore concluded that seismic shaking will not cause reactivations.

- Groundwater is an important factor in landslide stability, and confined aquifers are the most unfavourable groundwater conditions. This is because water can build up next to clay layers and push the slide downhill, or water can be trapped beneath the slide base so high destabilising pressures can build up over a large area. During lake-filling reactivation was quite likely, because the toes of many of the landslides were submerged. In a dry landslide, this can soften dry clay minerals causing a loss of strength. It can also affect groundwater systems, and water decreases the effective weight of the rock and so decreases the frictional resistance in the toe.

CHAPTER THREE

Landslides in the Northwest Part of the Wakatipu Basin

3.1 Introduction

In this chapter the aspects of the Coronet Peak and Arthurs Point Landslide that will be discussed are;

- 1) Their extent (area, volume, location) and nature.
- 2) What mechanisms they failed by, whether they are kinematically feasible or not and what the likely failure surface was.
- 3) Their present stability, including rates of movement.

3.2 Investigation Techniques

3.2.1 *Aerial Photograph Interpretation*

Aerial photographs were obtained from New Zealand Aerial Mapping Land Information Services, Hastings, for the purposes of interpreting stereo pairs to identify features on the ground surface. Enlargements of a photo covering some of the study area were obtained at a scale of 1:10 000. This photo is number 2823/12, survey number 1219, flown on the 18 February 1959.

3.2.2 Engineering Geological Mapping

The aims of the field mapping were:

- 1) To determine the extent of the Coronet Peak Landslide and other landslides in the Northwest region of Wakatipu Basin.
- 2) To map these landslides in detail and to determine failure mechanisms and geomorphological development models for these landslides
- 3) To assess the suitability of the landslides, in particular Coronet Peak and Arthurs Point Landslides for residential development.

3.2.3 Field Investigation

The only field investigation undertaken was augering some holes in sag pond areas on the Coronet Peak Landslide and on glacial surfaces (Fig 1.6a-map pocket) to look for material suitable for carbon dating. No suitable material was found, so no dating of the landslide or glacial surfaces was possible. A log of what material was found in these holes is given in Appendix Three.

3.3 Arthurs Point Landslide

3.3.1 Extent and General Description

Arthurs Point Landslide is located on the southwest slope of Bowen Peak, and is a wedge failure (Fig 3.1). The toe reaches the Shotover River, but the slide does not extend all the way to the ridge crest of Bowen Peak. It has an area of 0.34km^2 and an estimated volume of $2.4 \times 10^7\text{m}^3$.

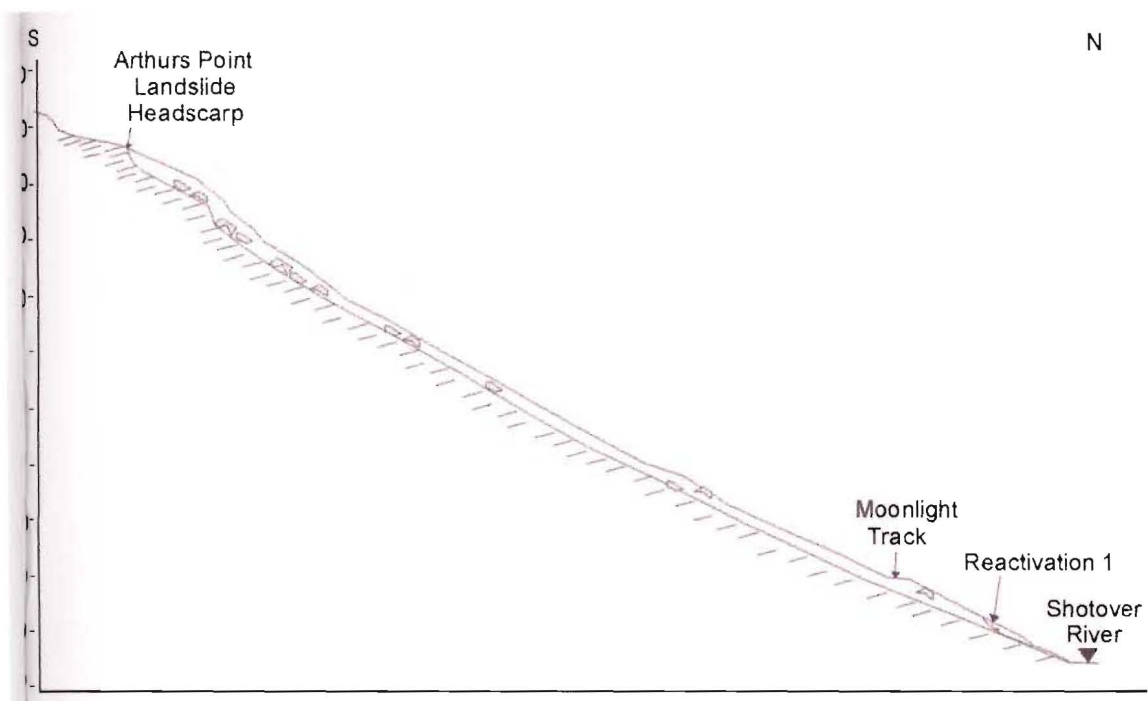


Fig 3.1 Cross-section of Arthurs Point Landslide.

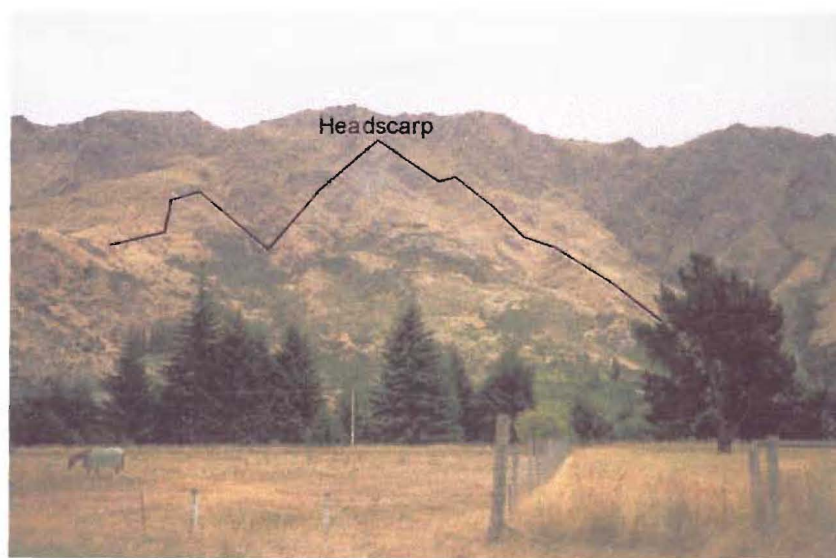


Fig 3.2 Photo showing Arthurs Point Landslides prominent head and lateral scarps.

With an average slope angle of 40° (ranging from 30° to 45°) the Arthurs Point Landslide is very steep. Elevation ranges from 1360m at the headscarp to 460m at the toe, and the vertical depth perpendicular to the failure plane is estimated to be from 20m to 65m (Fig 3.1).

Being a wedge failure, the landslide is characterized more by intact schist blocks that have moved downslope and prominent scarps than ripple-landscape. The term 'ripple-landscape' refers to the terrain that forms by slow deep-seated creep of the slide downslope, over the lifetime of the slide. This terrain typically forms on steeper slopes ($22^{\circ} +$), and shows large areas of subdued, apparently stable, landslide topography surrounding smaller areas of reactivation.

Ripple-landscape does occur, so the slide must be experiencing a certain amount of creep, but this terrain is not well developed. Topography is steeper at the toe and head, with flatter areas in between. The prominent joint-controlled headscarp shows a definite v-shape, which is characteristic of wedge failures (Fig 3.2; Fig 1.6a-map pocket). The lateral scarp to the west is also prominent and joint-controlled; while to the east the lateral scarp is prominent only from the headscarp down to an elevation of 760m (Fig 3.2; Fig 1.6a-map pocket).

It is inferred that glacial over-steepening caused initial failure, and then the slide continued to fail retrogressively. Two areas of reactivation are present (Fig 3.3; Fig 3.4; Fig 1.6a-map pocket), associated with incision of the Shotover River into the toe of the landslide (possibly about 5000 years ago or less).

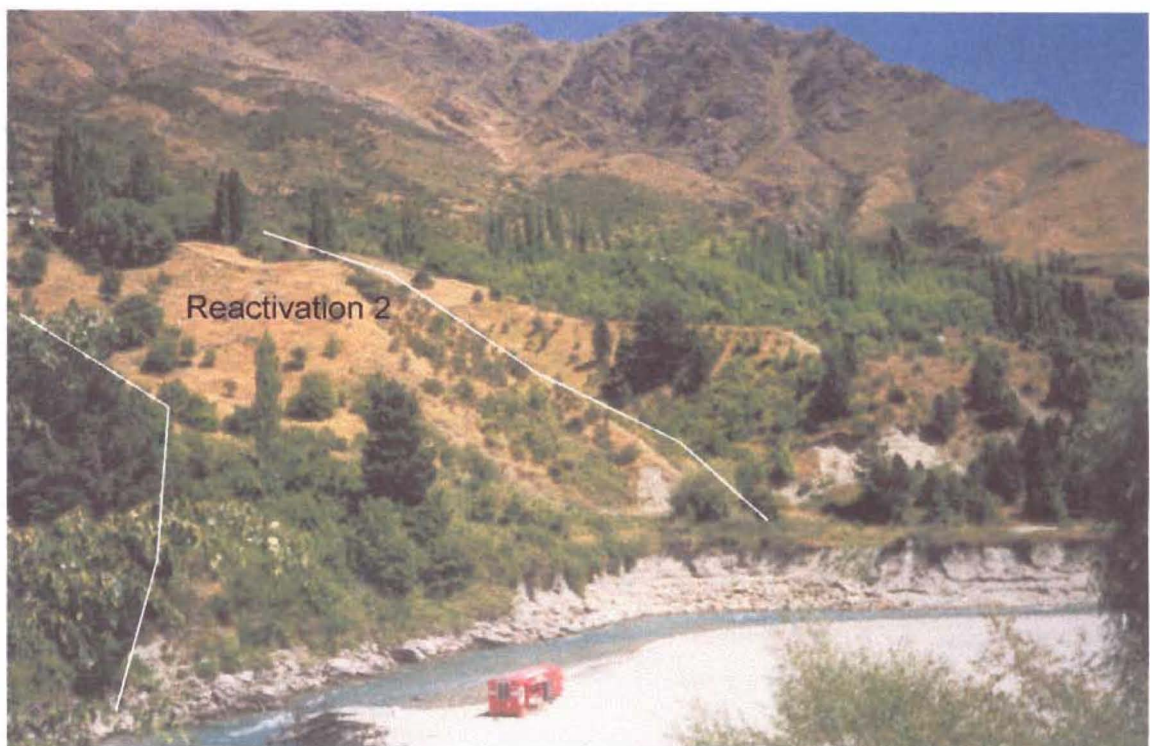
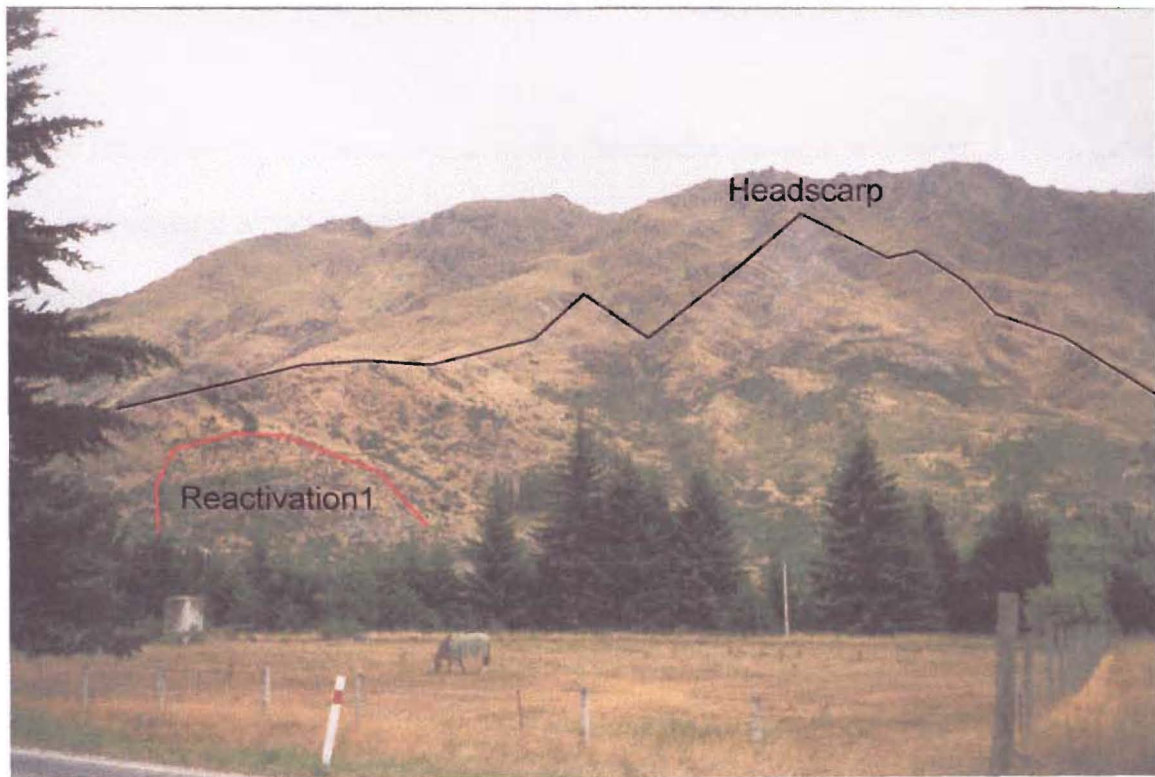


Fig 3.4 Toe of reactivated part of Arthurs Point Landslide.

No large seepage zones occur in this landslide, though one stream does cross it, and several small seepages can be seen along the Moonlight Track. In this context 'small' is defined as a dampness seeping out of the batter face, approximately affecting a 15cm wide area. No actual stream is formed from this seepage. Swampy areas are present and are associated with the stream, which is reasonably well incised (Fig 1.6a-map pocket).

The Arthurs Point Landslide has developed in the Otago Schist. Formation of the schist has been discussed in detail in Chapter One, and therefore will not be covered here.

Only psammitic and pelitic greyschist lithologies are present in the landslide area. The psammitic schist is described as unweathered to moderately weathered, very strong to moderately strong, light brownish grey to dark grey, coarsely foliated quartzofeldspathic schist. Pelitic schist is unweathered to slightly weathered, very strong to weak, dark grey, foliated, micaceous schist. All rocks in the area belong to textural zone IV.

Characteristics of the rockmass (defects such as foliation, schistosity, joints, faults, crushed zones and foliation shears) have developed as a result of the schists history of metamorphic and tectonic activity.

Landslide deposits can be described as chaotic debris, which consists of schist blocks that are both block and matrix supported, with the matrix consisting of silty sandy gravel.

A 0.3m-0.5m thick cover of schist colluvium overlies this. The engineering geological description of the colluvium is slightly weathered, dry, loose, massive, brownish grey schist colluvium. A 250mm to 300mm thick topsoil horizon is present over this, which indicates a significant period of time.

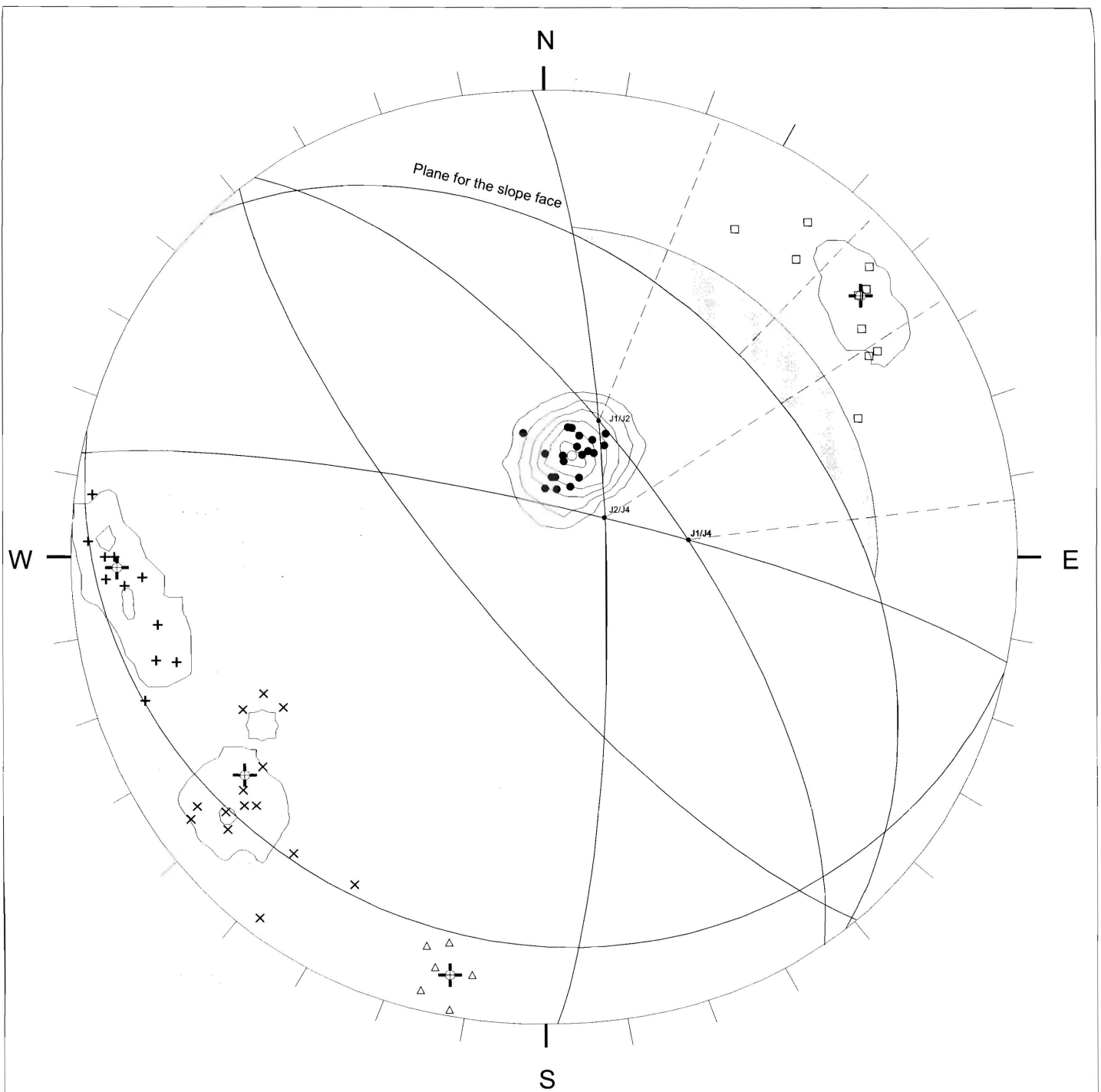
3.3.2 Failure Mechanisms

With an average foliation attitude of $284/18^{\circ}\text{S}$ (varying from 298° to 270° on the strike), the foliation dips into the slope face, and therefore cannot be the surface upon which sliding occurred (Fig 3.2). Instead two joint sets intersect to form a plane along which sliding occurred. Four main joint sets are present: $J1=320/70^{\circ}\text{NE}$, $J2=180/81^{\circ}\text{E}$, $J3=325/69^{\circ}\text{SW}$ and $J4=287/80^{\circ}\text{N}$.

The controls on the structure that must be satisfied for wedge failure to occur are; following Turner and Schuster (1996): pg 403-404

- 1) The trend of the line of intersection must approximate the dip direction for the slope face.
- 2) The plunge of the line of intersection must be less than the dip of the slope face. Under this condition, the line is said to daylight in the slope.
- 3) The plunge of the line of intersection must be greater than the angle of friction of the surface.

Three intersections of joint sets forming possible failure planes can be seen in Fig 3.5. The trends of the intersections are; $J1/J2=022$, $J2/J4=057$ and $J1/J4=084$.



KEY

- Poles For Foliation
- × Poles For Joint Set 1
- + Poles For Joint Set 2
- Poles For Joint Set 3
- △ Poles For Joint Set 4
- Poles For Planes
- Friction Angle (30°)

Fig 3.5 Stereographic Projection of poles and planes to foliation, joints and the slope face. In this figure the slope face is at an angle of 40°

As the trend of the slope face is 044, all of the joint set intersections approximate the dip direction of the slope face. Therefore, the first criterion for wedge failure is satisfied.

The plunges of the lines of intersection are more than the dip of the slope face (40°) so failure could not occur, as they do not daylight in the slope (Fig 3.5). If the slope had a steeper angle (like 70° - 80°) then the intersection lines would daylight in the slope. Steepening of the slope face during glacial advance and retreat so that the slope face was near vertical, or incision into the bedrock by the Shotover River are the likely mechanisms. It is worth noting that the present Shotover Gorge has very steep sides, with slope angles ranging from 70° - 75° . If an angle of 75° was used, then the plunge of the lines of intersection would be less than the slope face, and the second control would be satisfied (Fig 3.6).

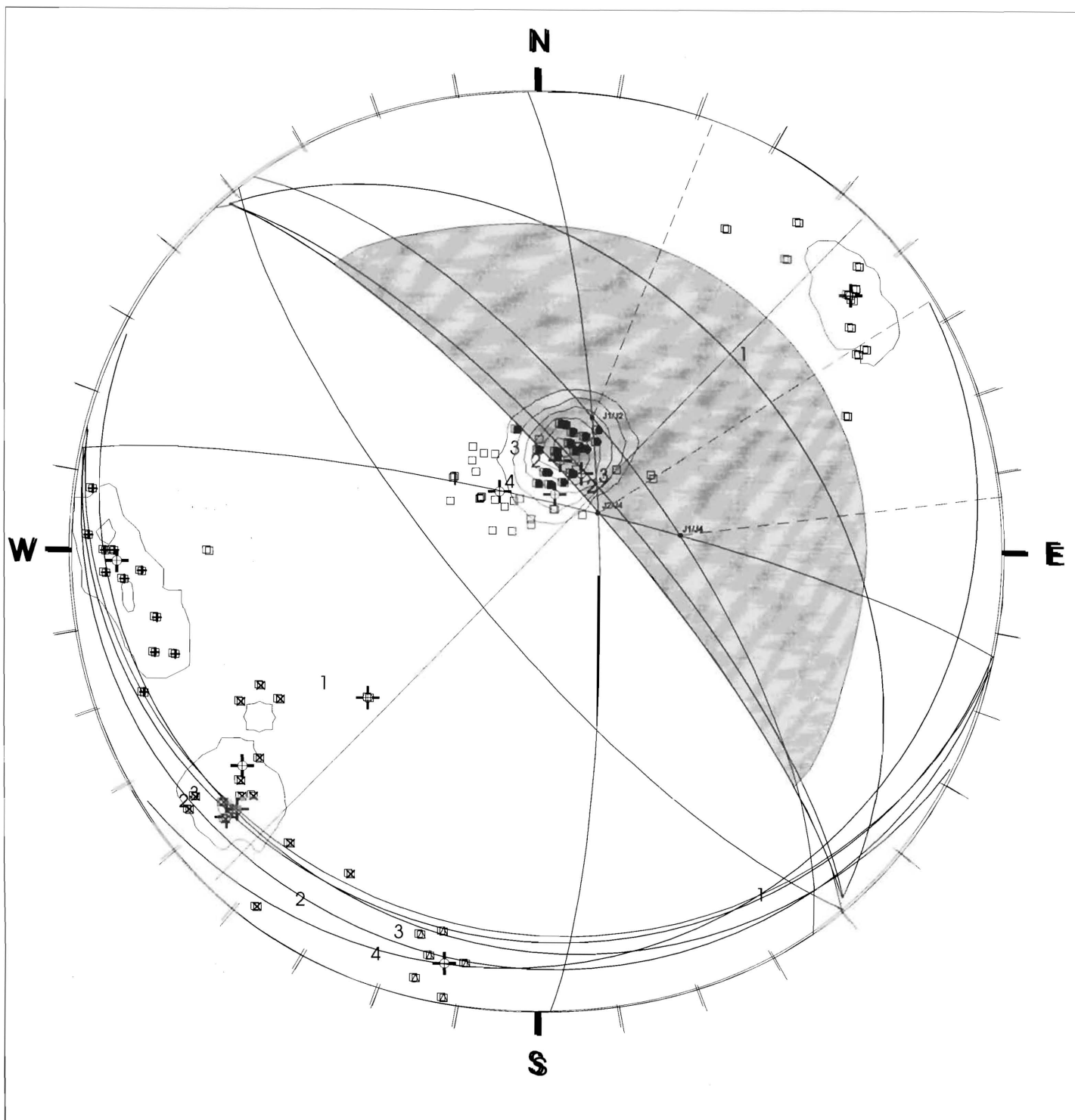
If we assume that the slope face was at an angle of 75° then the friction angle would have to be exceptionally high for sliding not to have occurred, so the third control is satisfied as well (Fig 3.6).

Subsequent down cutting of the Shotover River to form the gorge below the Arthurs Point settlement caused two reactivations (Bell 1995e).

Since then river terraces and other fluvial deposits have modified the toe and removed material so that a normal toe bulge is not present (Fig 3.1).

3.3.4 *Present Stability*

Due to the lack of hummocky terrain, and the reasonably well-incised stream, it is inferred that the majority of the slide is creeping very slowly, and is therefore quite stable.



KEY

- Poles For Foliation
- × Poles For Joint Set 1
- + Poles For Joint Set 2
- Poles For Joint Set 3
- △ Poles For Joint Set 4
- Poles For Planes
- Friction Angle (30°)

Fig 3.6 Stereographic Projection of poles and planes to foliation, joints and the slope face. In this figure the slope face is at an angle of 75°.

Two areas of reactivation are present, associated with the down cutting of the Shotover River. It is likely that these areas are moving at a faster rate than the rest of the slide as the support for them will be continually undermined.

3.4 Devils Creek Landslide

3.4.1 Extent and General Description

Devils Creek Landslide is located on the next ridge northwest from the southern end of the Coronet Peak Landslide, with Devils Creek running along its toe (Fig 1.6a-map pocket). Running laterally from Mt Dewar to the Shotover River it extends from ridge crest to valley floor. It has an area of 3.5km^2 and the volume is estimated to be $4.6 \times 10^8\text{m}^3$. Elevations range from 1163m to 380m, and the average slope angle is 24° . The estimated vertical depth perpendicular to the slide base is 80m.

3.4.2 Failure Mechanisms

It is likely that the Devils Creek Landslide is a foliation-controlled translational failure, as it is situated on a similar ridge to Coronet Peak Landslide. The ridges run the same way, and the landslides are situated on the same side of the ridges, with similar foliation attitudes (Fig 1.6a map-pocket). Therefore it is assumed that the Devils Creek Landslide is kinematically feasible.

It is likely that failure occurred along a combination of foliation and defects sub-parallel to foliation.

3.4.3 *Present Stability*

Devils Creek Landslide has very well developed hummocky topography, indicating that the slide has definitely experienced some creep movement. It is inferred that the rate of this movement is around 8mm/yr or less.

Devils Creek itself runs along the toe of Devils Creek Landslide, and acts as a form of toe support removal. Several reactivations are present (Fig 1.6a-map pocket). Three are situated along Devils Creek and are related to possible flood events and undermining of the slope. Two other reactivations are present, related to the incision of the Shotover River into bedrock.

3.5 **Coronet Peak Landslide**

3.5.1 *Extent and General Description*

The Coronet Peak Landslide is a schist block and debris slide located on the northwestern side of the Wakatipu Basin. Extending from ridge crest to valley floor it has an area of 23km² and an estimated volume of 1*10⁹m³.

Elevation ranges from a maximum of 1651m (Coronet Peak itself) to 450m at the valley floor, and the range of slope angles are from 11° to 26°. Slope angles are steeper in the head and toe regions of the landslide, while the middle is gentler. The vertical depth perpendicular to the slide base is a maximum of 150m.

The Coronet Peak Landslide can clearly be divided into zones, based on surface morphology. Zone A is a larger zone, which has more subdued morphology. A prominent lateral scarp defines the boundary to the east, while a prominent headscarp is present along the ridge up to Coronet Peak itself (Fig 1.6 map pocket).

Zone B is clearly reactivated, indicated by the abundance of scarps present. Zone B is divided into three sub-zones, based on age of movement. The zones are marked on the map (Fig 1.6a+b-map pocket).

3.5.2 *Zone A*

3.5.2.1 Extent and General Description

Zone A of the Coronet Peak Landslide extends from the NE lateral scarp (Fig 1.6a–map pocket) through Coronet Peak itself, through Skippers Saddle up to Mt Dewar and finishes at the edge of the reactivated area (Fig 1.6a-map pocket).

Its area is 19km^2 and the estimated volume is $6.0 \times 10^8\text{m}^3$. Steeper slopes occur in upper areas and in the toe, with gentler slopes in between. Ripple landscape is the most abundant terrain although it is very subdued.

3.5.2.2 Failure Mechanisms

The Coronet Peak Landslide is a translational failure in bedrock, controlled by foliation and/or defects parallel to foliation. The average foliation attitude of insitu schist measured near the slide is $232/15^\circ\text{SE}$, while the dominant joint sets are J1 at $323/75^\circ\text{SW}$ and J2 at $300/83^\circ\text{NE}$.

All three lithologies of schist are present in Zone A. The description for greenschist is unweathered to slightly weathered, very strong to strong, light green, finely foliated greenschist. In psammitic schist the joint sets are spaced from 15cm to half a metre, while in pelitic schist they are less common and more closely spaced. Instead crushed zones and chevron folds are more likely in pelitic schist.

A crushed zone, or foliation shear (approximately 100mm or less), outcrops in situ pelitic schist (Fig 3.7) located on the Devils Creek side of the saddle that is immediately south of Mt Dewar (Fig 1.6a-map pocket). It is inferred from this evidence that there are many more foliation shears present in the slide and bedrock, and many chevron folds have been observed. No faults strike through the slide, but the Shotover Fault strikes very near it at Queenstown Hill.

Several structural factors have to be met in order for it to be possible for planar failure to occur and these are; following Turner and Schuster (1996): pg 395-396

- 1) The dip direction of the planar discontinuity must be within 20° of the dip direction of the slope face (or the strike of the planar discontinuity must be within 20° of the strike of the slope face).
- 2) The dip of the planar discontinuity must be less than the dip of the slope face and thereby must "daylight" in the slope face.
- 3) The dip of the planar discontinuity must be greater than the angle of friction for the surface.

- 4) The lateral extent of the potential failure mass must be defined either by lateral release surfaces that do not contribute to the stability of the mass or by the presence of a convex slope shape that is intersected by the planar discontinuity.

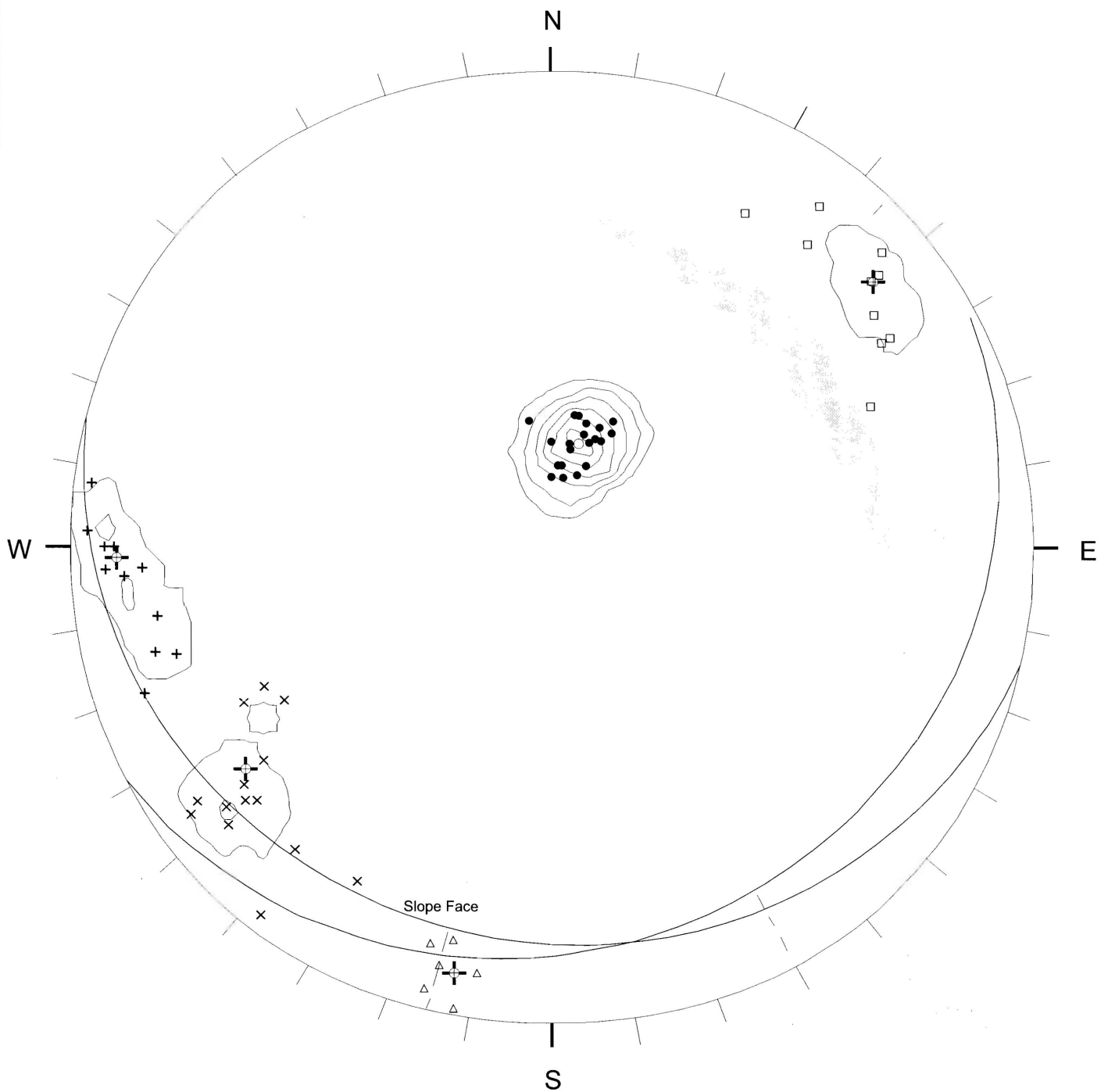
At Coronet Peak Landslide there is a swing in strike of the foliation. In the part of Coronet Peak Landslide that heads northeast from Mt Dewar (Fig 1.6a-map pocket) the foliation strikes at an average of 235° while the strike of the slope face is approximately 244° .

For the part of Coronet Peak Landslide that heads south from Mt Dewar the average strike of the foliation is 220° and the slope is oriented at about 210° . This gives differences of 9° and 10° respectively. So even though the strike of the foliation swings, the orientation of the slope face changes too, meaning that the first criterion for planar failure is always met (Fig 3.8).

The next factor is that the dip of the planar discontinuity (15°) is less than the slope angle (25°), allowing the discontinuity to daylight in the slope, so this is also satisfied. For sliding to occur the frictional element of the failure surface has to be overcome, so the dip of the discontinuity must be greater than the friction angle. As sliding has already occurred this criterion has definitely been met. In pelitic schist friction angles can be quite low (10° - 15°), and they can be even lower in foliation shear zones (the probable failure surfaces).



Fig 3.7 Photo showing foliation shear zone located near Coronet Peak Landslide.



KEY

- Poles For Foliation
- × Poles For Joint Set 1
- + Poles For Joint Set 2
- Poles For Joint Set 3
- △ Poles For Joint Set 4
- Poles For Planes
- ▣ Friction Angle (30°)

Fig 3.8 Stereographic Projection of poles and planes to foliation and the slope face. In this figure the slope face is at an angle of 0° .

Lastly, release surfaces are needed to allow the slide to separate from the main rockmass, and joint sets identified in the rockmass provide these releases.

A combination of foliation shears and crushed zones make up the probable Basal Failure Zone. A crushed zone outcrops in the field area, so it is inferred that other crushed zones and foliation shear zones exist.

Sliding was initiated during the retreat of a lobe of the Wakatipu Glacier at the end of the Waimean Glaciation (135 000yrs). During the Waimean (Penultimate) Glaciation the northern lobe of the Wakatipu Glacier extended for some distance down the Kawarau River, and subsidiary ice tongues pushed up the Arrow and Shotover valleys (Fig 1.5).

3.5.2.4 Present Stability

Zone A of the Coronet Peak Landslide is relatively stable. No big areas of reactivation are present. Streams are deeply incised, indicating no large-scale movement has occurred since the initial failure. There is abundant hummocky terrain (Fig 3.9), implying ongoing movement of the slide in the form of downslope creep.

This probably at a slower rate than seen at Cromwell Gorge, as there is no form of toe support removal (e.g. a river, another landslide, a glacier). At Coronet Peak Skifield there has been no noticeable movement of any of the pillars that support the skilift over a time period of 50 years.



Fig 3.9 Hummocky topography of the Coronet Peak Landslide

3.5.3 *Zone B*

3.5.3.1 Extent and General Description

Zone B refers to the section of Coronet Peak Landslide from the Shotover River to just before Downeys Dam, from ridge crest to valley floor. The area is 3.3km² and the area is estimated to be 4.0*10⁸m³.

3.5.3.2 Failure Mechanisms

There is no change in failure mechanisms from Zone A to Zone B. It is still a translational foliation controlled failure, which satisfies all the criteria for planar failure. Foliation shear zones and crushed zones form the Basal Failure Zone, and initiation of the slide is retreat of the Wakatipu Glacier. Several reactivations are present.

3.5.3.3 Present Stability

This is the more active part of Coronet Peak Landslide. One area of reactivation is associated by further glacial activity during the Otiran advance. Another area is related to incision of the Shotover River, and the youngest area of reactivation is associated with flood events on the Shotover River.

It is probable that this zone is creeping at a similar movement rates to the Cromwell Gorge slides.

3.6 Other Landslide Features

3.6.1 *Dirty Four Creek Slide*

The Dirty Four Creek Slide is located along the ridge to the west of Coronet Peak (Fig 1.6a-map pocket), and has an estimated volume of $1 \times 10^8 \text{ m}^3$. It has hummocky topography, indicating that it is fairly stable.

It is likely to be a translational failure on foliation or defects parallel to foliation. It appears to be fairly deep-seated.

From the topography it would appear that the slide is moving slowly, probably at rates less than 10mm/yr, as it is only undermined at the toe by a stream.

3.7 Comparisons With Previous Mapping

3.7.1 *Barrell et al, 1994.*

3.7.1.1 Landslide Features

Barrell et al (1994) has mapped many landslide features, but not in detail. No geomorphic information (such as headscarps, lateral scarps, secondary failures) or structural data has been provided (3.10).

Specific differences between the author's map and Barrells' map do exist. On Barrells' map the toe of the Arthurs Point Landslide does not extend all the way down to the Shotover River, whereas the author mapped the toe as being at river level (Fig 1.6a-map pocket). Actual field reconnaissance by the author showed that the toe of the landslide is being modified by the Shotover River, as well as there being no geomorphic evidence for a toe occurring further upslope.

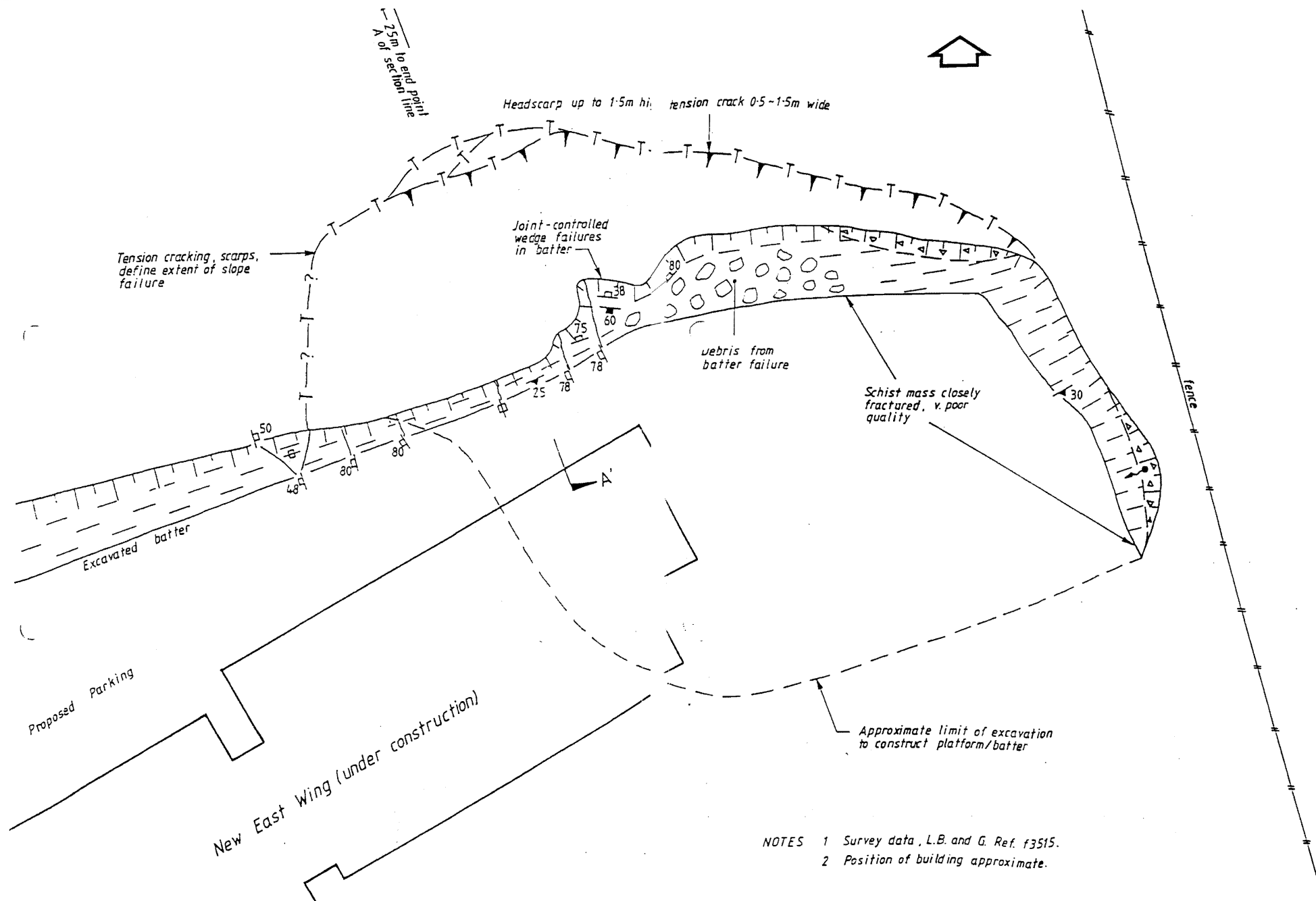


Fig 4.3 Plan Of Coronet Plaza Wedge Failure

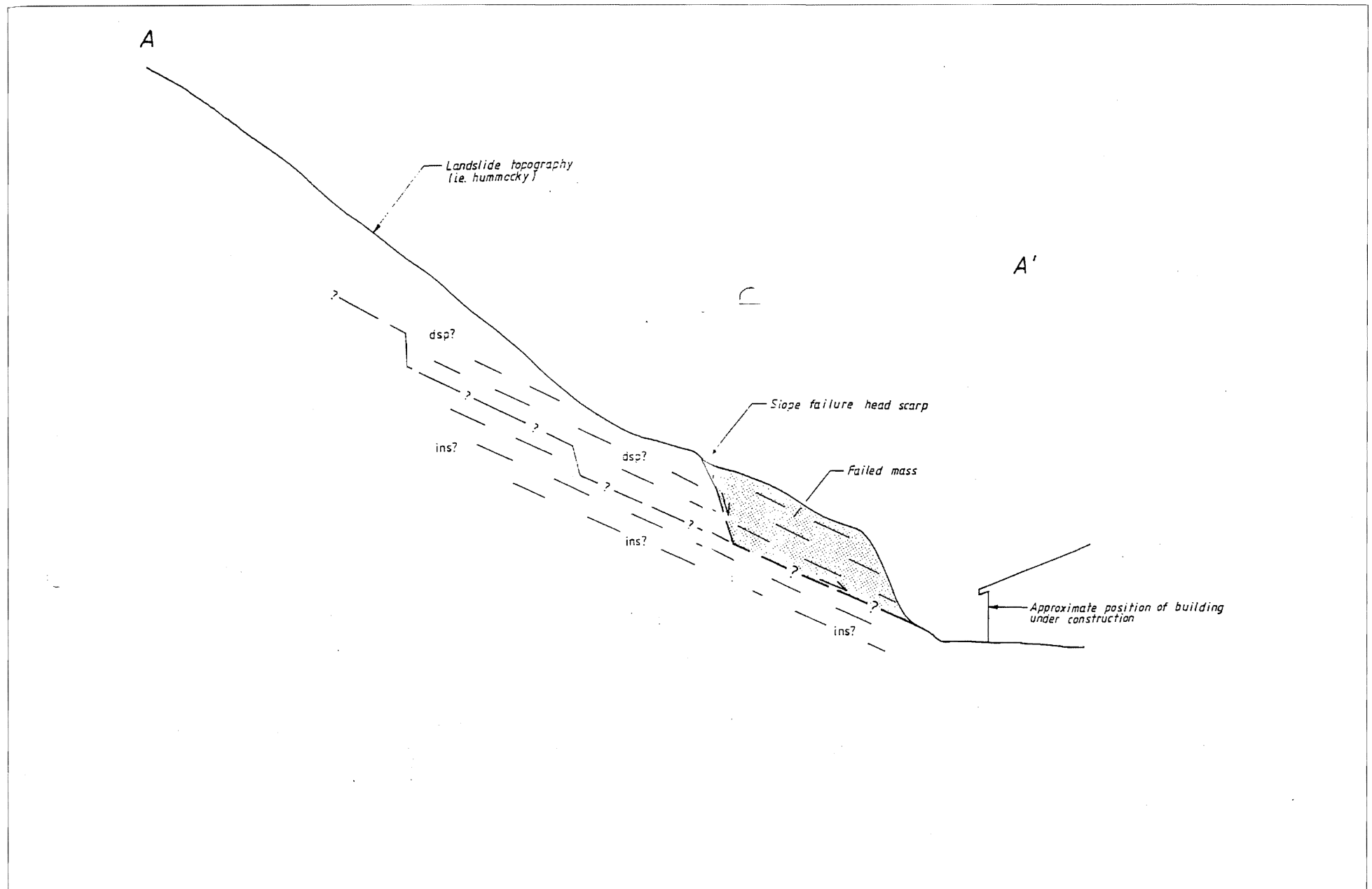


Fig 4.2 Cross-section of Coronet Plaza wedge failure.

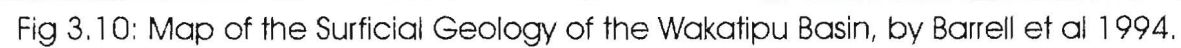


Fig 3.10: Map of the Surficial Geology of the Wakatipu Basin, by Barrell et al 1994.

Aerial photographic analysis by the author shows both lateral scarps extending down into the Shotover River.

For the Coronet Peak Landslide there are several differences between the mapping of Barrell et al (1994) and the author. Most of these differences relate to separate slides being included in the Coronet Peak Landslide.

A small slide is present just past the southern lateral scarp of the Coronet Peak Landslide. Barrell et al (1994) have included this slide in the Coronet Peak Landslide, while the author has mapped it as a separate slide (Fig 1.6a-map pocket). This slide is clearly separate on aerial photographs, and field evidence also shows a clearly separate slide with it's own headscarp and lateral scarps.

A second slide is present directly to the NE from Mt Dewar, which again has been mapped as part of the Coronet Peak Landslide by Barrell et al (1994). From the first aerial photograph analysis; the author initially mapped this slide as part of the Coronet Peak Landslide. As a result of fieldwork and the construction of a cross-section through Mt Dewar (Fig 1.7) the author decided that the slide was separate from Coronet Peak Landslide.

The last slide that Barrell et al (1994) mapped as part of the Coronet Peak Landslide is the Dirty Four Creek Landslide, which is mapped separately by the author (Fig 1.6a-map pocket).

3.7.1 2 Glacial Features

Barrell et al (1994) mapped the glacial deposits as an undifferentiated unit (moraine and till). Outwash terraces are included in a unit called 'Terrace Alluvium' which also includes a variety of river-laid deposits. In contrast, the author has mapped the glacial deposits of till, moraine and outwash as separate units (Fig 1.6a-map pocket).

3.7.2 *Cunningham (1994)*

As with Barrell et al (1994), Cunningham (1994) has also mapped the surficial deposits of the Wakatipu Basin, but has also mapped geomorphology and geological structure.

3.7.2.1 Landslide Features

With respect to the Arthurs Point Landslide, Cunningham (1994) agrees with the author that the toe of the landslide extends all the way down into the Shotover River.

Cunningham (1994) also agrees with the author concerning the small slide to the west of the southern lateral scarp of the Coronet Peak Landslide, as she has mapped it as a separate slide.

The second slide has been mapped as part of Coronet Peak Landslide by Cunningham (1994) as has the Dirty Four Creek Slide.

While the author has mapped the northwestern lateral scarp just over a kilometer west from Coronet Peak itself (Fig 1.6a-map pocket), Cunningham (1994) has mapped the lateral scarp as occurring further along the ridge (almost at Arrowtown). The author mapped the lateral scarp closer to Coronet Peak because of field observation.

3.7.2.2 Glacial Features

Cunningham (1994) has included glacial geomorphic features on her map and, like Barrell et al (1994), has chosen to map the glacial deposits in one unit. This differs from Barrell et al (1994) in that she has included outwash terraces in the unit.

3.8 Synthesis

- Arthurs Point Landslide is a large failure ($2.4 \times 10^7 \text{ m}^3$), with prominent joint-controlled scarps.
- Failure of the Arthurs Point Landslide occurred on the intersection of two joint sets (J1 at $300/70^\circ \text{NE}$ and J2 at $180/81^\circ \text{E}$), when the slope was oversteepened, either by glacial or fluvial incision.
- The age of the Arthurs Point Landslide is inferred to be related to the last (Otiran) glaciation.
- Arthurs Point Landslide is thought to be a retrogressive failure, with failure occurring in the SE part first, and then retrogressive failure of the rest due the rock above being unsupported after the initial failure.
- Two reactivations in the toe are present, related to incision of the Shotover River, moving at faster rates than the rest of the slide.

- Devils Creek Landslide is a translational failure, most likely on foliation or defects parallel to foliation. This slide is fairly stable, inferred by the well-developed hummocky terrain present. It is thought that the slide is moving at rates slower than 10mm/yr.
- Coronet Peak is a large-scale failure, occurring in bedrock, and can be divided into zones based on morphological evidence.
- Zone A is a fairly stable part of Coronet Peak Landslide. Failure has occurred along most likely a foliation shear zone, with lateral release on steeply dipping joints. Movements rates are quite slow (less than a few mm/yr)
- Zone B is a more active part of Coronet Peak Landslide, and can be divided into three subzones. There is no change in failure mechanisms across zones. However, movement rates will be faster, due to the presence of a toe failure support mechanism.
- As well as the Coronet Peak and Arthurs Point Landslides there are other landslides present in the immediate area around them. Most of the landslides around Coronet Peak Landslide are translational failures, forming on dipslopes, with similar foliation attitudes.
- Barrell et al (1994) mapped the surficial deposits of the Wakatipu Basin, but did not map types of landslides or morphological features. The main differences between their map and the authors is that they have included landslides that are not part of the Coronet Peak Landslide as part of the landslide complex.

- Cunningham (18994) mapped the same area, and again, the main differences were the inclusion of slides that were not part of the Coronet Peak Landslide in the landslide complex.

CHAPTER FOUR

Coronet Peak Landslide Movement History

4.1 Introduction

In this chapter the following aspects of Coronet Peak Landslide that are discussed are;

- The cause of initial failure of the landslide and the likely triggering mechanism.
- What age the initial failure is, looking at geomorphological evidence.
- What the ages of the three reactivations are, and likely triggering mechanisms.
- The Coronet Plaza failure and the geotechnical aspects involved.

4.2 Initial Failure

4.2.1 *Geological Setting*

During the Miocene, tectonic activity associated with the Kaikoura Orogeny has reshaped the Central Otago landscape. Block faulting and limited folding of the bedrock due to regional compression has transformed the once flat surface ('peneplain') into multiple elevated ranges and structural depressions. The ridge along which Coronet Peak Landslide has formed is part of one of these ranges, while the Wakatipu Basin is one of the depressions.

Slope development is strongly influenced by the nature of the schist bedrock as foliation provides continuous planes of weakness. In the case of the ridge on which the Coronet Peak Landslide formed, the foliation is dipping into the valley at an angle of 15° plus or minus 5° , while the slope face is at about 20° (Fig 1.7). Other defects that are sub-parallel or parallel to the foliation (such as schistosity, foliation shears and crushed zones) also add to the general instability of the slope.

These defects are unfavourably orientated and are ready to fail when some form of slope oversteepening occurs, such as glacial or fluvial incision. When either the glacier retreats or the river incises to a depth sufficient to cause gravitational spreading, failure can occur due to the slopes being left unsupported.

4.2.1 *Back Analysis*

The back analysis was performed as a function of the Galena program. Assumptions made by the program are that the failure surface is planar and that the minimum cohesion is zero. Input of certain parameters is necessary; the material that is being analyzed, the minimum ϕ value (which in this case was 10°) and the factor of safety required (a value of 1.00 was used for equilibrium, and 0.800 was used to ensure the slope to fail).

During back analysis the strength parameters of all other materials in the slope are kept constant, while combinations of the cohesion and angle of shearing resistance of the nominated material are determined which produce the required factor of safety.

Five Coronet Peak Landslide cross-sections were analyzed; XSA, XSC, XSD, XSE and XSF. The results of the back analysis are summarized in Table 4.1.

Apart from XSD the maximum ϕ for a factor of safety of 1.00 ranged from 20°-22°, while for a factor of safety of 0.80 the maximum ϕ was 15°-18°.

From an extensive review of the literature on testing of the schist (Atkins and Bryant, 1977; Paterson, 1979; Watts, 1988 and many others summarized in Appendix 2) average values have been found. The average value of the friction angle assumed for intact schist is 28°, which suggests failure is very unlikely to occur through it. Crushed schist and fractured schist have an average friction angle of 20°, making failure through that kind of schist again unlikely. This indicates that failure most likely occurred along a foliation shear zone, for which an average friction angle of 10° has been assumed.

4.2.3 *Triggering Mechanism*

It is inferred from geomorphological evidence that the initial failure of Coronet Peak Landslide was triggered by glacial incision. Fluvial incision is another possible triggering mechanism, but is thought not to have had an influence in this case. The Shotover River runs past the southern lateral scarp of Coronet Peak Landslide (Fig 1.6a-map pocket), and even may have run along the front of the landslide as far as the end of Zone B1 (although no evidence for this has been seen in the field), but it is unlikely that it flowed in front of the landslide the whole way along the toe.

TABLE 4.1 Summary of Back Analysis Results

	Factor of Safety	Maximum Phi (o)	Maximum Cohesion kPa
XSA	1	21	200
	0.8	17	130
XSC	1	21	150
	0.8	16	75
XSD	1	16	165
	0.8	13	75
XSE	1	20	187
	0.8	15	23
XSF	1	22	200
	0.8	18	100

Also, other landslides where fluvial incision may have played a part (K9 Landslide in Kawarau Gorge, most of the Cromwell Gorge landslides and the Gibbston Slide in Gibbston Valley) are all formed where the river has incised into the bedrock to form a gorge, and the Shotover River has not incised a gorge anywhere along the front of Coronet Peak Landslide.

4.3.3 Age Of Failure

It is known that at least four glaciations have occurred in the Wakatipu Basin in the last 500 000yrs (already discussed in Chapter One), and it is therefore reasonable to conclude glacial incision has to be the triggering mechanism. It is also known that during these glaciations the northern lobe of the Wakatipu Glacier carved out bedrock to form the Wakatipu Basin and the valley in front of Coronet Peak Landslide. During the largest (or Penultimate) glaciation (the Waimean) ice extended down the Oreti and Mararoa valleys, down the Kawarau Valley to Waitiri, over the Crown Terrace (600m above sea level) and up the Arrow River (Fig 1.5). In the Otiran (Last) glaciation the ice did not reach as far (Fig 1.5), depositing terminal moraine (Old Ben Lomond Station moraine) in the bedrock valley adjacent to the northern boundary of Zone B1 (Fig 1.6a-map pocket). Ages of the Waimean and Otiran glaciations are 135 000yrs ago and 26 000yrs ago respectively.

As ice did not extend all the way along the front of Coronet Peak Landslide in the Otiran, then it is unlikely that it was that glaciation that triggered the landslide, and therefore the slide cannot be younger than 26 000yrs (the Gibbston advance). Instead, the likely glaciation to have triggered the Coronet Peak Landslide is the Waimean, which constrains the age of the slide to slightly younger than 135 000yrs.

4.3 Zone A

Zone A is the largest zone of Coronet Peak Landslide (Fig 1.6a-map pocket), and due to the lack of secondary scarps, is inferred to be fairly stable.

Investigations of the Cromwell Gorge landslides gave values of 10mm/yr as an average rate of creep movement. Movement rates and direction of movement could be investigated for the Coronet Peak Landslide. However, the amount of time needed to get a significant result would be more than the time allowed to complete a Masters Thesis. Therefore, no investigation of rates of movement of the Coronet Peak Landslide has been undertaken.

After initial failure it is unlikely that the Coronet Peak Landslide was moving at the rates seen in the Cromwell Gorge. At the Cromwell Gorge most of the landslides have their toes periodically removed by the Clutha River, whereas at Coronet Peak Landslide there is no form of toe removal (such as a river, glacier or another landslide). Deposition of glacial deposits and outwash terraces during the Otiran glaciation would have overlain the toe, and therefore had a buttressing effect, reducing the rate of movement even more.

Streams that are present in this part of the landslide are well incised and at the northwestern end have formed very large coalescing alluvial fans that almost spread right across the valley floor (Fig 1.6a-map pocket). Such large alluvial fans, which are still centered on the streams, would imply that no significant movement has occurred in the slope since initial failure, though it would still be creeping at a rate of a few mms a year.

4.4 Zone B

4.4.1 Zone B1

Zone B1 is the largest subzone of the reactivated area of Coronet Peak Landslide (Fig 1.6a-map-pocket). Since the initial failure, associated with the Waimean glaciation (approx. 135 000yrs ago), it is likely that Zone B1 has been creeping at a very slow rate (less than 10mm/yr) due to the lack of any mechanism of continued toe support removal.

From the geomorphological evidence (Fig 1.6a-map pocket) it is obvious that this part of the landslide has undergone more downslope movement than Zone A. This is suggested by the many prominent scarplets throughout the slope, which are relatively fresh compared to the subdued hummocky landscape present in Zone A. A prominent secondary scarp inside a more degraded headscarp would tend to indicate that this movement has occurred more recently than the initial failure.

The age of this reactivation can be constrained by the evidence of two geomorphic factors. The first is that the terminal moraine mapped (Fig 1.6a-map pocket) in the valley, referred to by the author as Old Ben Lomond Station moraine, and the northern edge of Zone B1 coincide quite nicely. As the Old Ben Lomond Station moraine was deposited by the Wakatipu Glacier during the Otiran glaciation, this would indicate that the glacier eroded out the sides of the bedrock valley up to the limits that it reached (Fig 1.5), also providing a mechanism of toe support removal. When the glacier retreated it left the valley sides oversteepened and unsupported, which led to the reactivation.

The second piece of geomorphic evidence is that it can be inferred from aerial photograph analysis that the toe of the slide mass in places overlies alluvial and glacial terraces (Fig 1.6a-map pocket). No exposures have been seen in the field to confirm this as there were no funds for trenching, but it can also be noted that places where this occurs are associated with areas of reactivation. Therefore, places where the toe overlies alluvial and glacial terraces must mean that the reactivations are younger than the last glaciation (15 to 25 000yrs ago), but not the initial movement.

In Zone B1, the direction of movement also gives a clue about what triggered the reactivation, because the direction of movement is towards where the glacier would have been (Fig 1.6a-map pocket), whereas the other reactivated areas have a sense of movement towards the Shotover River.

4.4.2 *Zone B2*

As the second largest subzone of Zone B, Zone B1 has a sense of direction towards the Shotover River (Fig 1.6a-map pocket), which indicates that the Shotover River is the likely triggering mechanism for the reactivation. This would give the reactivation a probable age of less than 5000yrs old. Now covered by vegetation, it is nevertheless quite easy to pick out the reactivation boundaries (Fig 4.1).

4.4.3 *Zone B3*

Zone B3 is the youngest and smallest reactivation of Zone B (Fig 4.1), made up of two small slides in the toe of the slide at the southern end. The headscarp of one of the slides (the one closet to the main headscarp of Coronet Peak Landslide (Fig 1.6a-map pocket; Fig 4.1) is prominent and joint and foliation-controlled.

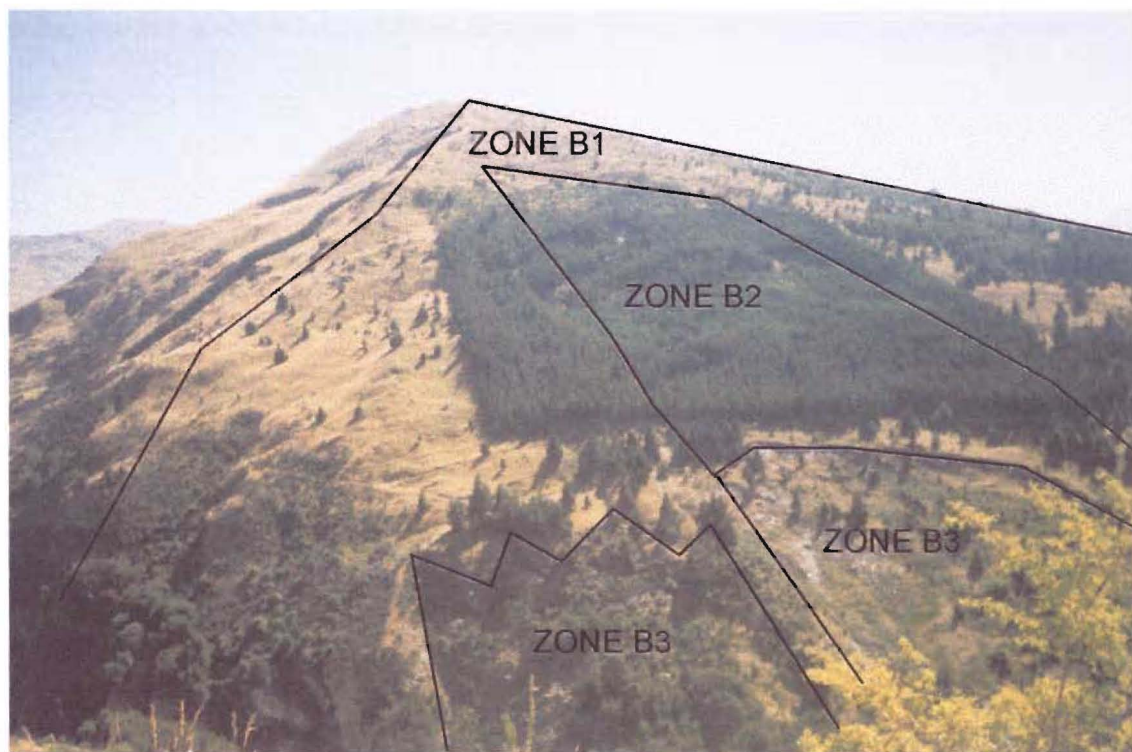


Fig 4.1 Zone B1, B2 and B3 of the Coronet Peak Landslide.

As the two slides have similar areas and volumes, it is inferred that they are related to the same flood event by the Shotover River. Since these areas of reactivation have developed in Zone B2, then they must be younger. Smaller reactivations are present within this zone (Fig 4.1), demonstrating the effectiveness of the Shotover River as a toe support removal mechanism. Each of these reactivations within reactivations are certainly related to many subsequent flood events of the Shotover River, each getting progressively younger.

4.5 Coronet Plaza Failure

The Coronet Plaza, formerly known as the Shotover Resort, is situated at the toe of the Coronet Peak Landslide just past Arthurs Point. The slope of the landslide toe initially rises at 25° , but steepens to an average of 40° (Fig 4.2). The building itself is on a flat glacial (outwash) terrace. In early March 1995, excavations were being carried out to enable construction of a new east wing (now built), which caused a slope failure to occur. The excavations were carried out to increase the available flat land on which to put the new building, along with vehicle access and parking.

The area of failed slope above the 10m steeply cut batter is delimited by surface tension cracking (0.5-1.5m wide) and scarps (up to 1.5m high) extending over a length of 60m (Riddolls and Grocott 1995). Fresh opening of persistent widely spaced joints perpendicular to the rock face provide evidence of movement in the batter itself (Fig 4.3), and a local joint-controlled wedge failure has occurred in the center.

The failed mass consists of schist blocks that have already been previously affected by landslide movement. This is indicated by infill in open joints and poor rock mass quality. The slope failure is inferred to comprise “displaced schist” overlying “undisturbed (insitu) schist” using a classification scheme developed by MacFarlane et al (1992) for local schist.

Thinly laminated and moderately-slightly weathered, the schist is very fissile. Foliation typically dips 25° to the south, and it is considered that slope failure occurred along a sliding plane sub-parallel to this as foliation was dipping out of the slope. Steeply dipping joints sub-parallel to the batter provides release surfaces for sliding to occur. The volume of the failed mass is estimated at 2000m^3 .

Although there was no information on slope movement prior to excavation, it is likely that ongoing slope movement contributed to the wedge failures that occurred. Failure occurred the day after excavation work was completed, so it is most likely that failure occurred due to removal of support of the toe of the slope.

On the 20th of March 1995, removal of locally potentially unstable rock blocks from part of the batter was carried out to reduce safety hazards while construction was in progress. Though movement on the slope may have slowed or ceased, there is still a risk it may move in the future. Prolonged rainfall, rapid snowmelt or even earthquake shaking can all cause reactivation of the slope failure.

Therefore, some remedial stabilization work is needed. Removal of unstable material from the head of the slide (approx. 3m) has improved stability by 7-8%, but there is still the need to assess the possibility of inducing additional instability further upslope caused by the excavation required.

Need to ensure further upslope regression does not occur by foliation-controlled sliding in the schist. Creation of a stable "key-block" above the headscarp can be achieved by rock bolting or buttressing, and a schist block wall could also be established at the toe of the slope to act as a catch bench and to provide access for clearing any debris that accumulates by fretting (Riddolls and Grocott 1995).

The Coronet Plaza failure is the youngest reactivation, and occurs within Zone B1. Triggering of this slide by injudicious excavation of the landslide debris indicates the toe of the Coronet Peak Landslide is in a very delicate equilibrium.

4.6 Synthesis

- Development of basin and range topography during the Kaikoura Orogeny (in the Miocene) uplifted the schist bedrock, which in some areas formed foliation dipslopes where the foliation dips into the valley at a moderate angle (eg 15°).
- The inherent weakness of the foliation and other defects oriented unfavorably (that is sub-parallel or parallel to the slope) contributed to the failure of the slope.
- Back analyses conducted indicated failure was unlikely to occur through intact rock (with a ϕ of 28°) or crushed and fractured schist (with a ϕ of 20°). The results showed that failure along a foliation shear zone, with a ϕ of 10° , is the most likely failure surface.
- Initial failure was most likely triggered by glacial incision, rather than fluvial incision, shown by evidence seen in the field.
- The age of initial failure has been constrained to slightly younger than the Waimean glaciation (approximately 135 000 yrs ago) and older than the Otiran (approximately 26 000yrs ago).
- Zone A is not reactivated and therefore is inferred to be of the same age as the initial failure. Movement rates for this zone of CPL are less than 10mm/yr.

- Zone B1 is part of the reactivated area of CPL, with abundant scarps. This zone will have been creeping at less than 10mm/yr before initiation of the reactivation. The age of the reactivation has been demonstrated by evidence seen in the feed to be associated with the Gibbston advance.
- Zone B2 reactivation was triggered by the Shotover River, which gives it a probable age of less than 5000 years old.
- Zone B3 is the smallest reactivation of Zone B, and reactivation was triggered possibly by a Shotover River flood event.
- The Coronet Plaza failure is the youngest reactivation event. Failure occurred due to injudicious excavation of the toe of the slope, and failure occurred as a wedge and steeply dipping joint sets provided release surfaces.

CHAPTER FIVE

Geomorphic History and Planning Implications

5.1 Introduction

In this chapter the items that are covered are:

- 1) A detailed glacial history of the Wakatipu Basin.
- 2) Identification of geotechnical constraints affecting the Coronet Peak and Arthurs Point Landslides.
- 3) A discussion of whether it is feasible to build on these landslides.

5.2 Glacial Events

During the Quaternary tectonic activity associated with the Kaikoura Orogeny has reshaped the Central Otago landscape. Block faulting and folding of the bedrock due to regional compression has transformed an initially flat surface (peneplain) into multiple elevated mountain ranges and structural depressions (basins).

Also known as basin and range country the basins were initially poorly integrated. The development of a better-integrated drainage system occurred as the uplift reached its Pleistocene climax.

Most of the tertiary sediments have been stripped off the ranges due to erosion, but some of these sediments are still preserved in some basins, mainly as inliers along faults. No Tertiary sediments are preserved in the Wakatipu Basin.

Folding and faulting are superimposed on a background of ongoing regional uplift. It is possible that the Queenstown-Arowtown basin and the valley occupied by Lake Wakatipu are downfolded areas, which have been cut into and modified by glacial action, but not principally formed by erosion.

Alternatively, these areas may have been uniformly uplifted during the Kaikoura Orogeny, and these basin and valley areas have been entirely formed by glacial and river erosion, leaving the surrounding ranges as remnants of uplifted land (Barrell et al 1994). It is not known which of these models is correct.

During the last 2.5Ma years or so, the global climate has undergone a series of fluctuations, resulting in periods of generally cold temperatures (glacials) separated by periods of relative warmth (interglacial). It is probable that at least 20 separate glacials have occurred during the 2.5Ma (Barrell et al 1994). There is evidence for only four of these periods of glacial advance and retreat, over a time period of 500 000 years.

During glacial periods, summer temperatures were very low. As winter snow did not melt from the main ranges, extensive snowfields accumulated in the high precipitation areas along the Southern Alps, from which large glaciers spread down the valleys. The glaciers have scoured out the valleys and basins as the ranges have risen. Permanent snowfields formed above about 1500 metres elevation on the highest ranges such as The Remarkables. This produced localized cirque glaciers that scoured out basins near the summit.

Ice brought down by the Wakatipu Glacier carved out the valley that is now Lake Wakatipu. Where the glacier met The Remarkables it split, one lobe heading northeast towards Arrowtown forming the main Wakatipu Basin ice tongue, while the main Wakatipu Valley ice tongue continued south towards Kingston. During some glacials the Wakatipu Basin ice tongue extended for some distance down the Kawarau Valley and a subsidiary ice tongue pushed inland up the Arrow and Shotover Valleys, while the Wakatipu Valley ice tongue has at times extended south beyond Athol in the Mataura Valley (Barrel et al, 1994).

Evidence for four main periods of ice advance and retreat has been identified in the Wakatipu Basin in the last 500 000yrs. During the oldest advance (Pre-Waitiri) the ancestral Shotover River was flowing north of Coronet Peak, depositing 150m of fluvial sediments in Deep Creek. The second oldest (and largest) advance known as the Waitiri and correlates to the Garston2 age deposits in the Wakatipu Basin. Next is the Kingston advance which is considered to be the Late Otiran Maximum, with an age of 25 000 years old. The Arthurs Point-Wye Creek occurred 18 000 year ago, where ice covered most of the Arrow Basin. It is considered by Turnbull and Forsyth, (1988) and Cunningham, (1994), to have deposited most of the glacial deposits in the Wakatipu Basin.

5.3 Coronet Peak And Arthurs Point Landslides

5.3.1 *Geotechnical Constraints*

When zoning areas of marginal land, the first thing to do is to identify the geotechnical constraints affecting that area, what parts they affect and the degree of limitation they impose. Three major geotechnical constraints affect the Coronet Peak and Arthurs Point landslide areas.

5.3.1.1 Landslip Hazards

Landslip hazards can be divided into two different categories, rockfall and shallow regolith failures. Rockfalls and topples are characteristic of steep back slopes controlled primarily by jointing, such as steep bluffs/slopes. Much of the footslope areas and outcrops of blocky debris have the potential to be affected by rockfalls, rolls and bounces.

It is extremely unlikely that any hazard is going to cause reactivation of either Arthurs Point or Coronet Peak Landslides. Any event that did reactivate them would have to be very large (with perhaps a probability of never happening within our lifetime).

However, reactivation of parts of either of these slides is very likely. Removal of support in an area, either by stream erosion or human excavation is one mechanism for reactivation. Another is a surcharge of weight from rain, snow (very likely on Coronet Peak itself), hail, artificial fill and buildings. Saturation of the slide mass as a result of heavy rainfall, stream diversion and artificial ponding of water can all possibly cause reactivation, as can seismic shaking.

These reactivations may take the form of isolated rockslides, rolls and falls of schist blocks and small debris flows, which have the potential to affect residential properties at the foot of the slope. Potentially very damaging to downslope property through deposition, erosion and flooding, small to large-scale reactivations may lead to landslide damming of streams, and subsequent breach and debris flow.

5.3.1.2 Flooding Hazards

Two physical processes associated with flooding can be identified; Storm events and catchment or channel modification.

During heavy rainfall events the degree of erosion increases significantly resulting in an increased sediment load. Poorly controlled storm water discharged into roadside gutters causes undercutting of roads. Melting of the snow at Coronet Peak Ski field results in a high flow of water down the drainage channels, which one year undercut Skippers Road and caused collapse of the edges of the seal.

Just as important to consider is aggradation, also known as channel building. Affected areas are stream channels and alluvial fan surfaces at the base of hills.

5.3.1.3 Seismic Hazards.

Both landslides are situated between two fault zones, the Moonlight Fault Zone and the Nevis-Cardrona Fault Zone. These faults have a probability of occurring in the next 150 years of 4 to 21%, and < 4% respectively (Gillon and Hancox, 1992). They both have similar recurrence intervals of about 4000 years.

It is considered unlikely that an earthquake will reactivate either of the CP or AP Landslides. This is because worldwide case histories suggest that the majority of landslides triggered by seismic shaking are first time failures, rather than reactivated pre-existing slides (Gillon and Hancox, 1992; Keefer, 1984).

5.4 Planning Implications

5.4.1 Coronet Peak Landslide

5.4.1.1 Coronet Peak Ski field Operations

Coronet Peak Ski field and its associated buildings have been on the Coronet Peak Landslide for fifty years, and haven't experienced any significant movement in that time, as no structural damage has occurred to any of the buildings or ski-lift pylons.

A few reactivations are present in the ski field area, but are only small, localized failures. They are likely to have occurred either under the weight of the snow, or due to increased water pressure in the ground when the snow melts.

No large-scale reactivations are present, implying very little movement is occurring in the form of creep. From studies done on the landslides in Cromwell Gorge, Kawarau Gorge and Gibbston Valley we know that most large-scale failures in schist usually experience some form of creep movement after initial failure. Averages for creep movement given in various publications are from 8-10mm/yr. If this part of the landslide is moving that much, then the slide should have moved 40-50cm, which is a significant amount of movement.

Also the thickness of the slide prevents the buildings foundations from penetrating the slide base into insitu rock. So either the buildings are moving along with the slide or the slide isn't moving much. It is inferred that a bit of both is happening. With no mechanism of ongoing toe support removal, it is unlikely that the slide is as fast as the average seen elsewhere. The slides from which this average was obtained all have a river flowing along the toe, which would continually undermine the slope above. It is therefore likely that the part of Coronet Peak Landslide that the ski field is built on is moving, but at a rate much slower than 8-10mm/yr.

5.4.1.2 Rural vs. Residential

In Zone A residential development is feasible, but not at the usual densities. As this part of the landslide is fairly stable, so long as the foundations are excavated carefully and the surface and groundwater are controlled.

Zone B is more active, and much more likely to fail, especially in the toe area. As has been shown in the Coronet Plaza failure, injudicious excavation of the toe materials can cause the toe to reactivate. When excavation of foundations occurs, it can unload the slope, and the toe area is especially sensitive to this.

5.4.1.3 Close Subdivision

Close subdivision cannot be the most acceptable way of developing the land. This means that a larger area is subjected to excavation for foundations, and that more water will be discharged into the slope. This is undesirable, because water quite often plays a part in making slopes unstable.

This is shown by the fact that the Shotover River is continually causing small reactivations within pre-existing larger ones.

5.4.2 *Arthurs Point Landslide*

5.4.2.1 Long Term Movement Rates

Knowledge of movement rates of landslides is important, as it is necessary to know whether the slide area is still creeping or dormant. Surveying regularly for more than several years is the only way to monitor with any certainty how much a slide is moving. No survey work was undertaken to assess movement rates, as the length of the study was not long enough to get anything conclusive. A lot of surveying has been done for other slides (the Cromwell Gorge slides and Gibbston Slide). This information gives rates of 6mm/yr to 20mm/yr for the Gibbston Slide (Johnson 1986) and 12mm/yr for the Cromwell Gorge slides (Gillon and Hancox 1992).

Rates of movement for the Arthurs Point Landslide are probably comparable with those found for the Cromwell Gorge slides, because the Shotover River is continually removing the toe of the landslide.

Two reactivated areas present are inferred to be moving faster than the rest of the slide. However, vegetation is well established, indicating no large-scale failures have occurred recently. Also there are residences already present on the southern reactivation, and no noticeable damage has occurred to any of them.

5.4.2.2 Construction Requirements

At the moment Arthurs Point Landslide is in a state of equilibrium, and so is fairly stable. To minimize the chance of reactivating the slide, careful excavation of the foundations must be undertaken.

It is recommended that foundation materials are heavily compacted, and large excavations are minimized. This poses limitations for drive-on access. Because of the nature of the foundation material there is likely to be variability in bearing capacity. Obviously block supported debris will have higher bearing capacity than matrix supported. Even differential settlement is possible unless careful compaction and drainage is carried out around the building site. Additional slab reinforcement, drainage of the foundation materials and long-term control of surface and subsurface water are necessary.

5.4.2.3 Housing Density

On landslides it is not recommended to build at usual densities. This is because residential development at its usual density would put too much weight on the slide material. Too much water would be discharged, and as water is an important factor in destabilizing a slope this could reactivate localized areas.

Excavation into the foundation material would be large-scale, and therefore increases the likelihood of reactivation.

5.4.2.4 Septic Disposal

It is imperative that appropriate location and design of the septic effluent disposal field is undertaken, to avoid compromising slope stability in any way. Preferably, the sewage tank should not be located on the slide mass, but if they are then the associated soakage trenches are very important. As it could reactivate the slide, the effluent cannot be left to soak into the landslide materials.

5.5 Synthesis

- The Wakatipu Basin has been extensively glaciated, though most of the landforms that are present relate to the Last (Otiran) Glaciation.
- The Coronet Peak Landslide can be used for residential development, but not at the usual densities because this increases the area that needs to be excavated, and would increase the amount of water needing to be controlled.
- Zone A is the better part of the landslide to build on, as it is moving at a considerably slow rate (less than 8mm/yr).
- The toe of the landslide would be the least desirable place to develop residential property, as this is a part of the landslide that is already prone to reactivation.
- Excavation of the foundations and control of surface and subsurface water would be the two most important things to take into account when building on them.

<p style="text-align: center;">CHAPTER SIX</p> <p style="text-align: center;">Summary and Conclusions</p>

1) Arthurs Point Landslide is a wedge failure, with sliding occurring along the intersection plane of two joint sets. Failure was initiated by glacial oversteepening and incision of the Shotover River. It is a retrogressive failure, with glacial oversteepening causing the initial failure of the eastern part of the slide, which left the above rock unsupported. This, in combination with the incision of the Shotover River cause the western part to fail.

2) Coronet Peak Landslide is a translational failure, which has slid along a failure plane made up of a combination of foliation and other defects parallel to foliation (most likely foliation shear zones, which are known to occur in the field area).

3) Coronet Peak Landslide can be zoned based on geomorphological differences, and has been divided into Zone A and ZoneB1, B2 and B3.

4) Zone A is the largest zone, and is moving fairly slowly (less than 8mm/yr) while the reactivated areas (Zone B1, B2 and B3) are most likely moving faster due to toe support removal by the Shotover River.

5) Initial failure of the Coronet Peak Landslide was caused by glacial oversteepening of the slope, so that when the glacier retreated the slope was left

unsupported. As the foliation dips into the valley along this ridge and foliation shear zones are present, then the slope would be free to fail.

6) The age of the initial failure of Coronet Peak Landslide was during the Waimean Glaciation, which constrains the age to slightly younger than 135 000yrs old

7) Zone B1 is a reactivation that is associated with the Gibbston advance, while Zone B2 and B3 are associated with the incision of the Shotover River (approximately 5000yrs ago).

8) Because of the relative stability of Arthurs Point Landslide, it is feasible to build on it so long as it is not at usual densities, the surface and groundwater are controlled and the foundations are carefully excavated.

9) It is also feasible to build on the Coronet Peak Landslide as most of it (Zone A) is moving slower than Arthurs Point Landslide. The same geotechnical constraints are present as at Arthurs Point Landslide.

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APPENDIX I

List Of Surveys and Aerial Photographs

LIST OF AERIAL PHOTOGRAPHS USED

Date Survey Flown : 31 March 1954

Survey No 842 : Photo No :	2292/70
	2292/71
	2292/72
	2292/73
	2292/74
	2292/75
	2292/76
	2292/77
	2292/78
	2292/79
	2292/80
	2293/32
	2293/33
	2293/34
	2293/35
	2293/36
	2293/37
	2293/38
	2293/39
	2293/40
	2293/41
	2293/42

Date Survey Flown : 6 November 1957

Survey No 884 : Photo No :	232473
	232474
	232475
	232476
	232477
	232478
	232479
	232480

Date Survey Flown : 18 February 1959

Survey No 1219 : Photo No :	2823/12
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APPENDIX II

Summary Of Geotechnical Data

SUMMARY OF GEOTECHNICAL DATA

I Atkins and Bryant (1977)

Atkins and Bryant did uniaxial testing and found values for Young's Modulus (E) of 1.03 to 9.73 GPa, Poisson's Ratio (ν) of 0.07-0.29, failure stress (σ_1) of 2.77-21.93 MPa, confining pressure (σ_3) of 0 MPa, normal stress (σ_n) of 0.25-5.15 MPa and shear stress (τ) of 0.79-9.30 MPa.

Triaxial testing gave values of $\sigma_1=7.56-45.63$ MPa, $\sigma_3=0.75-5.90$ MPa, $\sigma_n=2.74-16.26$ MPa and $\tau=2.90-18.54$.

Unconfined compressive strength ranged from 22-96 MPa, with shear strengths of $c=1$ MPa, $p=22-33^\circ$ for intact rock, $c=0$ MPa, $p=11-23^\circ$ for fractured rock.

A summary of rock density measurements is given in Table I.I.

II Paterson (1979)

Paterson obtained peak ϕ of $22-32^\circ$. Looked at gouge material, found it had a considerably lower residual friction angle. Also found that it contains less silt, sand and gravel, but 25x more clay, with a substantially higher water content.

Samples containing 60% clay and 10% sand had ϕ of $6.6-7.7^\circ$, while samples containing 29% sand and 10% gravel had ϕ of 14° .

Also field strength of sheared schist and slope debris was given to be c_r of 0 and ϕ_r of 28.5° .

Table I.I				
Lithotype	Field Strength	Estimated UCS (MPa)		
		50-200		
1 Greyschist (psammitic)	strong-very strong			
2 Greyschist (psammitic)	mod. strong-strong	25-100		
3 Greyschist (pelitic)	mod. weak-mod. strong	5-50		
4 Greenschist	strong-very strong	50-200		
5 Greenschist	strong	50-100		

III Salt (1983)

Salt analysed schist debris slopes, using a large database from both natural and cut slopes in Kawarau, Cromwell, Clyde and Lake Roxborough areas and used the resistance envelope method to calculate field strength of the slope debris.

It was found that at low to moderate normal stresses (<150 kPa) $c=12$ kPa and $\phi=36^\circ$, and that a slight reduction in strength occurs at higher normal stresses $c=34$ kPa and $\phi=29^\circ$.

Field strength of sheared schist was determined from analysis of average basal stresses for active failures in the Kawarau Valley, $c_r=0$ and $\phi_r=28.5^\circ$.

IV Moody (1985)

Moody did Point Load testing of the schist at Maniototo Irrigation Scheme, and the results are summarized in Table III.I. He also measured crushed zone densities and did Schmidt Hammer Rebound tests on schist (summarized in Table III.II and III.III respectively).

V Johnson (1986)

Johnson found at low to moderate normal stresses (<150 kPa) slope debris gives values of $c=12$ kPa and $\phi=36^\circ$ while at higher normal stresses it gives values of $c=34$ kPa, $\phi=29^\circ$. For crushed schist from the Gibbston Slide the values range from $c_r=0$ and $\phi_r=27^\circ$ to $c_r=0$ and $\phi_r=32^\circ$.

Table III.I							
Rock Material	Location	Test Orientation	No. of	Is (50)	Value Of	Strength	
Description		(to Foliation)	Samples	(MPa)	Anisotropy		
			Tested	SD R M			
SW, light brownish grey, planar foliated, quartzofeldspathic schist.	Quarry	Normal	26	2.08 1.8-9.0 5.0	2.27	v strong	
		Parallel	26	1.00 0.5-5.0 2.2		strong	
UW, light grey, planar foliated quartzofeldspathic schist.	Link Race	Normal	13	0.79 0.5-2.8 1.4	1.14	strong	
		Parallel	21	0.48 0.7-2.4 1.2		strong	
UW, light grey, planar foliated quartzofeldspathic schist.	Quarry	Normal	14	1.47 4.2-9.4 5.8	2.15	v strong	
		Parallel	12	1.01 1.5-4.9 2.7		strong	
SW, dark grey, foliated micaceous pelitic schist.	Upstream Race	Normal	33	1.59 1.1-7.2 5.0	3.1	v strong	
		Parallel	41	0.44 0.3-2.2 1.0		m strong	
MW, brownish grey planar foliated quartzofeldspathic schist.	Downstream Race	Normal	11	0.97 1.1-4.4 2.2	2.75	strong	
		Parallel	11	0.30 0.5-1.5 0.8		m strong	
MW, brownish grey, planar foliated quartzofeldspathic schist.	Link Race	Normal	13	0.74 0.9-3.2 1.8	2.57	strong	
		Parallel	20	0.23 0.3-1.2 0.7		m strong	
S-MW, dark brownish grey, irregularly foliated quartzofeldspathic schist.		Normal	19	0.75 0.8-3.2 2.0	2.22	strong	
		Parallel	22	0.49 0.3-2.5 0.9		m strong	

Table III.II						
Description	Drillhole	Average Dry	Average S	Average	Average Particle	
Of Samples		Density	Density (kg/m3)		Density (kg/m3)	
		(kg/m3)				
UW-SW, strong,	DH1	2610	2650		2750	
brownish grey	6 Samples					
irregularly foliated						
quartzofeldspathic	DH5	2680	2700		2740	
schist.	4 Samples					
	DH9	2650	2680		2730	
	3 Samples					
	DH10	2610	2660		2750	
	2 Samples					
Rock Density measurements were obtained during investigations at the Middle Weir Site approximately 1.5km upstream of the Pateroa Power House. Rock was sampled from drillholes.						

Table III.III								
Rock	Rock Mass	Test	Test Orien-	No. Of	Rebound Number			
Material	Decription	Location	tation	Test Pits	SD	R	M	
Description								
UW, strong	Schistosity planar, dipping							
grey foliated	15-25oE; joints widely	Upstream	Perpen-	25	6.7	21-45	31	
pelitic schist	spaced, highly persistent,	Race	dicular					
	irregular, sub-vertical; no							
	groundwater (g/w) present.							
UW, mod.	Schistosity planar, dipping							
strong, grey	15-25oE; joints extremely	Upstream	Parallel	25	6.6	16-35	27	
foliated	widely spaced, highly	Race						
pelitic schist	persistent, irregular, sub-							
	vertical; no g/w present.							
UW, weak,	Schistosity dipping E;							
dark grey,	schist extremely shattered,							
foliated,	average block size 50-100	Upstream	Perpen-	5	-	-	0*	
micaceous	mm, contains many thin	Race	dicular					
pelitic schist	crushed zones (<5mm							
	wide); very low g/w flow.							
SW, strong,	Joints widely spaced,							
dark brownish	planar, slightly rough,	Link	Variable to	23	5.5	32-53	42	
grey, irreg-	highly persistent, medium	Race	Schistosity					
ularly foliated	sized unit blocks, no g/w.							
quartzofelds-								
pathic schist								
SW, strong,	Joints very widely spaced,							
dark grey,	planar, slightly rough,	Link	Variable to	13	5.8	22-43	33	
irregularly	highly persistent, large	Race	Schistosity					
foliated quartz-	sized unit blocks; no g/w.							
ofeldspathic								
schist								
UW, very	Joints very widely spaced,							
strong, planar	highly persistent, large	Quarry	Perpen-	5	1.7	50-54	52	
foliated	sized unit blocks, no g/w.		dicular					
quartzofelds-								
pathic schist								
* Rock Material too weak to obtain any readings.								

Gouge from the Muddy Creek and Gibbston slides has strengths of $c_r=0$ and $\phi_r=22^\circ$ to $c_r=0$ and $\phi_r=28^\circ$ at low normal stress (< 150 kPa) and strengths of $c_r=0$ and $\phi_r=19^\circ$ to $c_r=0$ and $\phi_r=25^\circ$ at higher normal stress (> 150 kPa).

Table V.I gives field rock strengths for the different lithotypes of schist.

VI MacFarlane (1984)

Point Load testing of drillhole core from KW 35 - KW 29 canal line showed that point load strength perpendicular to schistosity is 2.4x greater than point load strength sub-parallel.

VII Beetham et al (1992)

Strength and modulus values were found by testing intact schist, and they show that values are dependant on foliation orientation which typically lie around 5-105 Mpa (UCS) with an average of 40 Mpa. Also found that the average static modulus of compression parallel to foliation (av 30 GPa) was twice the value perpendicular to foliation.

VIII Jennings et al (1992)

Used back analysis to determine average shear strength of sheared schist.

The average back analysed value was $\phi=24^\circ$.

Direct and ring shear test results on gouge material removed from surfaces indicated little cohesion ($c=0$, $\phi=15-19^\circ$). Relative stability was calculated, and they found that slopes will be destabilised by the filling of Lake Dunstan, and that engineering works can offset the effects.

VIII MacFarlane et al (1992)

		Table VI				
Description Of Samples	Drillhole	Average Dry Density (kg/m3)	Average S Density (kg/m3)	Average	Average Particle Density (kg/m3)	
UW-SW, strong, brownish grey irregularly foliated quartzofeldspathic schist.	DH1 6 Samples	2610	2650		2750	
	DH5 4 Samples	2680	2700		2740	
	DH9 3 Samples	2650	2680		2730	
	DH10 2 Samples	2610	2660		2750	
Rock Density measurements were obtained during investigations at the Middle Weir Site approximately 1.5km upstream of the Pateroa Power House. Rock was sampled from drillholes.						

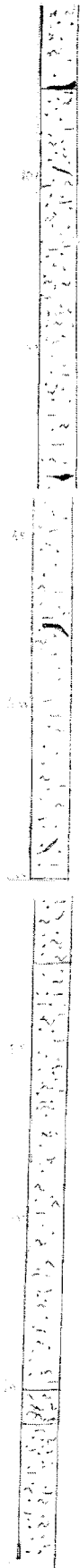
They found that mica-rich schists are in the weak to moderately strong category (q_u 10-50 MPa). Quartz-rich schists were found to be strong (q_u 30-90 Mpa), while massive and poorly laminate schists and greenschists were found to be very strong (q_u 100-250 Mpa).

Peak effective strength parameters parallel to foliation for quartz-rich schists were found to give $c=2$ Mpa and $\phi=23-32^\circ$. Effective strength of gouge material from the slide base determined using the resistance envelope method gave values for ϕ of $21-29^\circ$.

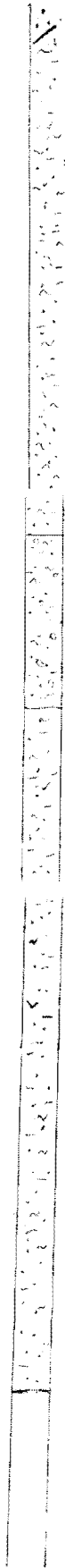
APPENDIX III

Sag Pond Logs

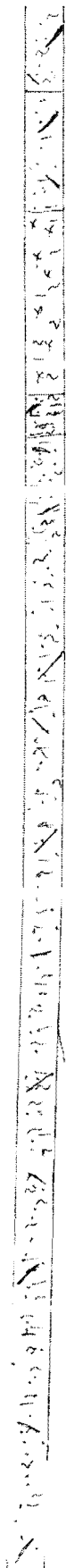
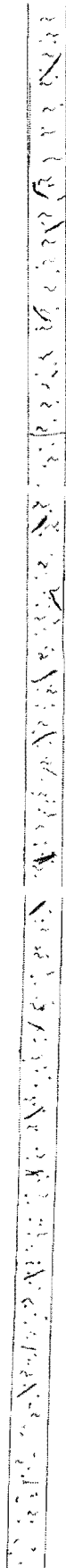
Ha1



Ha2



Ha3



KEY	
—	Clay
••	Charcoal
°°	Gravel
++	Sand
xx	Peat
~	Organics
∴	Silt
√	Roots

Ha4

Handwritten text in a cursive script, likely a form or document, spanning multiple lines.

Ha5

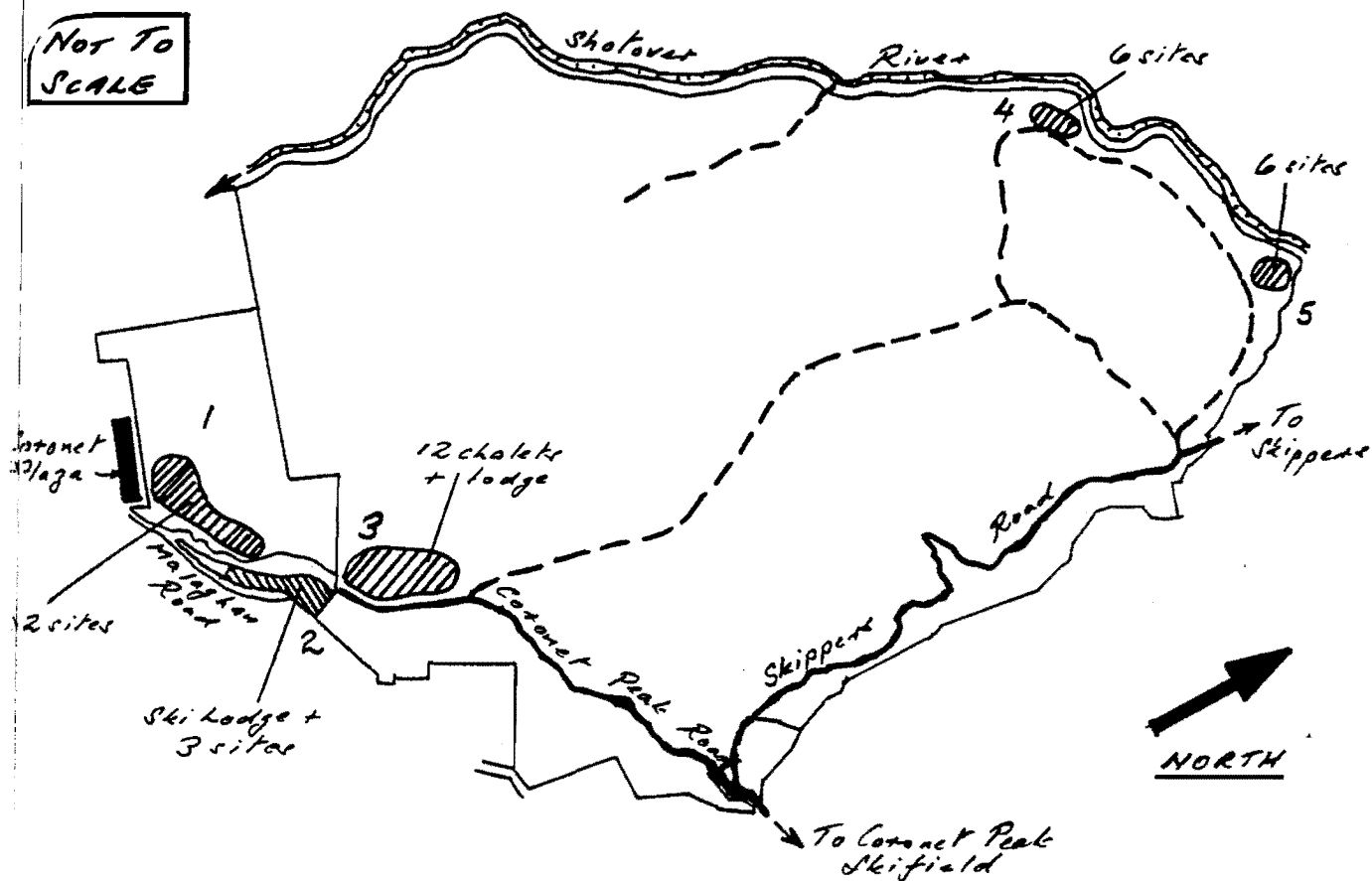
Handwritten text in a cursive script, likely a form or document, spanning multiple lines.

Ha6

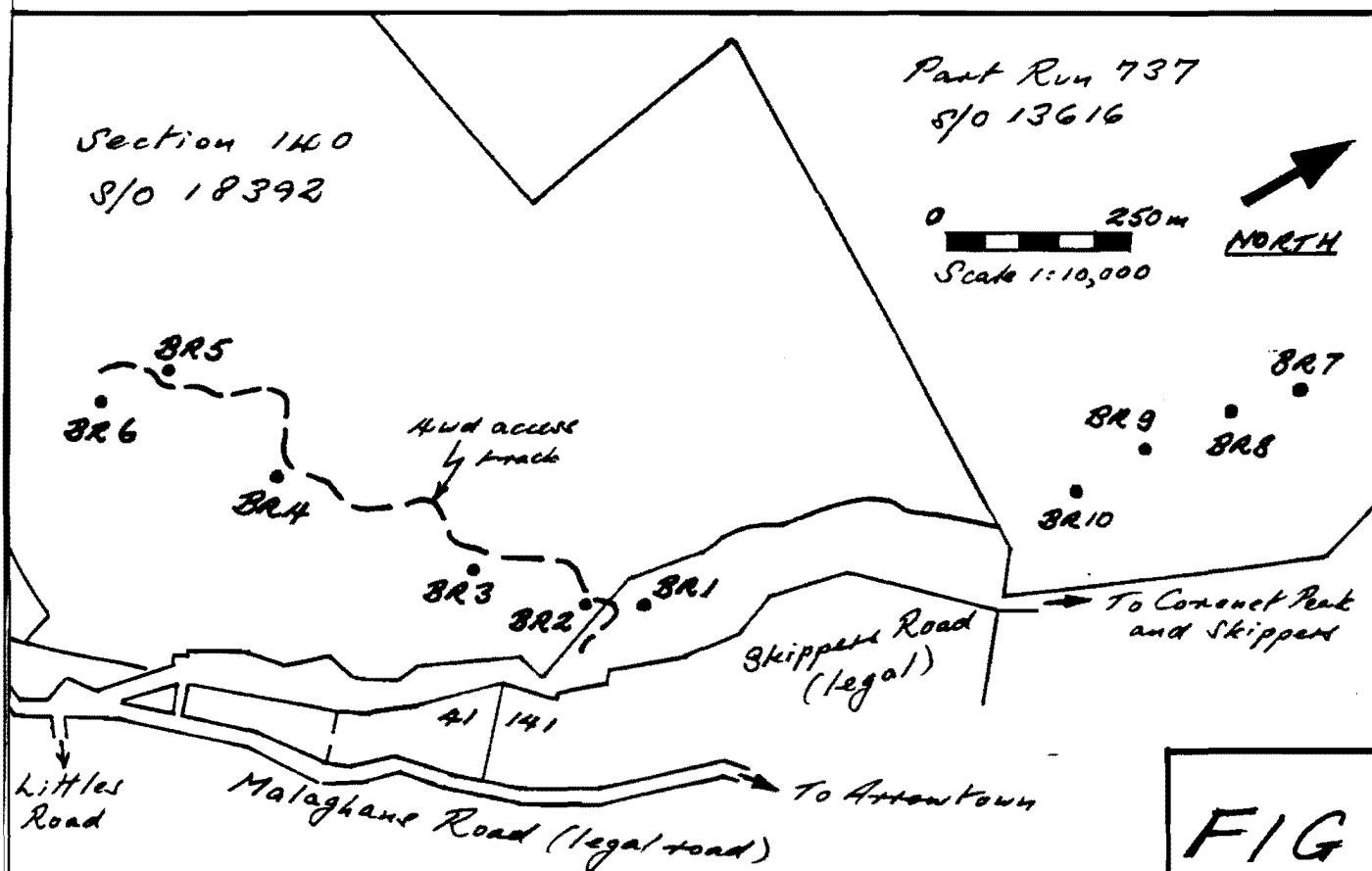
Handwritten text in a cursive script, likely a form or document, spanning multiple lines.

APPENDIX IV

Sample Logs and Locations



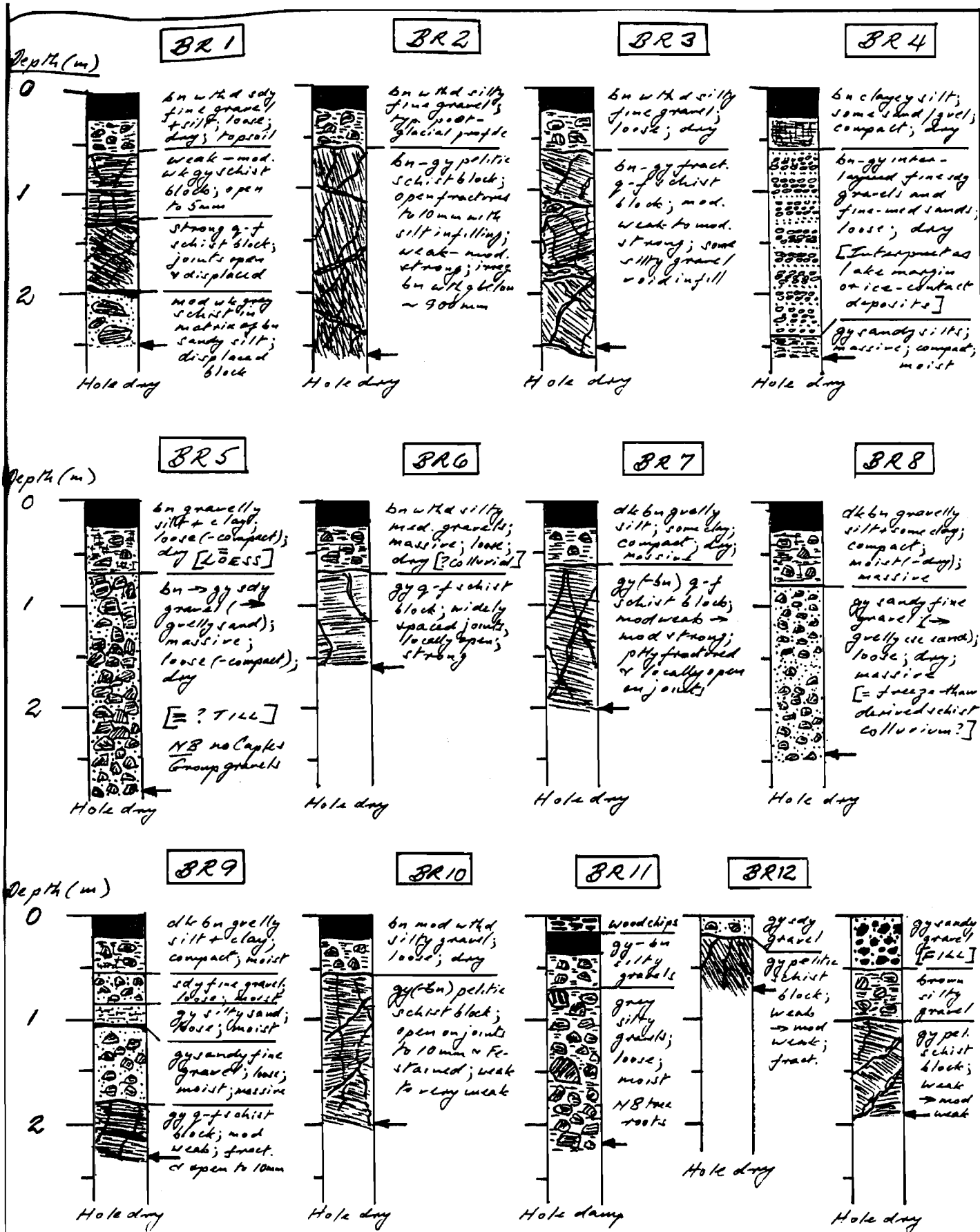
A. PROPOSED RURAL RESIDENTIAL SUBDIVISION



B. LOCATION OF TEST PITS 1-10 (Mt Dewar)
[Data from Clark Fortune McDonald & Associates]

FIG

1



Logged: DH Bell

19 or 20/2/96

Plotted: DHB 05/96

Surveyed: Clark Fortune
McDonald & Ass.

SUMMARY EXCAVATION

LOGS

Mt Dewar Property - Queenstown

(see Fig 1 for locations)

Scale:
1:50

FIG
2

MAP POCKET

Contains : Figure 1.6a and 1.6b

Figure 1.7a, 1.7b, 1.7c

FIGURE 1.6a

**Engineering Geological Map
at a Scale of 1 : 10 000**

FIGURE 1.6b

**Engineering Geological Map
at a Scale of 1 : 10 000**